

# Volcanic Seismology

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Fourth Edition

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# Preface to the Fourth Edition



I was glad to receive a letter from the Elsevier's Acquisitions Editor for the geoscience and atmospheric science Jennette McClain with invitation to prepare the Fourth Edition of my book. Now I prepared it, adding a new chapter about seismic events occurring during lava dome growth and destruction at andesitic and dacitic volcanoes, and four sections about the seismic events associated with significant volcanic events of the last decade. Among them are the Plinian submarine eruption of andesitic Hunga Tonga-Hunga Ha'apai Volcano in mid-January 2022, the first Plinian violent volcanic eruption in the past more than 100 years; then, the 2018 eruption of basaltic Kilauea volcano, Hawaii, including both a summit caldera collapse and a flank fissure eruption. It was the largest Lower East Rift Zone eruption and caldera collapse in at least 200 years. Finally, I added the data about seismicity of the 2020–21 eruptive episode at andesitic La Soufrière Volcano in the Lesser Antilles. Initial activity was an effusive phase that lasted three months and

turned explosive on April 9, 2021. I also added the description of the acoustic events, recorded during the 2022 Hunga Tonga-Hunga Ha’apai Volcano eruption. After consulting with the Elsevier officials, the title of this fourth edition of the book was changed from “Introduction to Volcanic Seismology” to “Volcanic Seismology”.

**Vyacheslav M. Zobin**

# Preface to the Third Edition

This August I celebrate the 50th anniversary of my life in *Volcanic Seismology*, and I am happy that the publishing house *Elsevier B.V.* had invited me to publish now the Third Edition of my book. For this edition, I have added a chapter about the seismic signals produced by the geothermal structures within volcanic environments. They also may be considered a component of volcanic seismology. The submarine eruption of new basaltic El Hierro volcano at the Canary Islands in 2011 gave a new experience for volcanologists, and I added its seismological description in Chapter 5. I partly rewrote the chapters about volcanic tremors and the seismic signals associated with volcanic explosions and volcanic flows. Other chapters were slightly reconsidered also.

**Vyacheslav M. Zobin**

# Preface to the Second Edition

The first edition of my book was published 8 years ago. I am glad to recognize that it became a textbook for many universities. Giving my lectures in *Volcanic Seismology* to graduate students of our Colima University and at some universities of the world, I understood that my book is needed in many improvements, especially in the part of theoretical basis. Many new results appeared in the study of eruption earthquakes. The continuous eruption at our Volcán de Colima, producing four lava extrusions during the period 1998–2011, gave invaluable collection of the thousands of seismic records of volcanic explosions, pyroclastic flows, and lahars, as well as simultaneous video records. Analysis of this database essentially served for my better understanding of the nature of volcanic earthquakes and their relationship with eruption process.

This book contains the data about volcanic earthquakes obtained during about 100 years. The quality and completeness of these data are different for different periods of observations. However, I preferred to include these characteristics of volcanic earthquakes because they were unique and not repetitive. Once I got the editorial comments to my paper where the editor recommended me to rewrite the manuscript using the seismic records of larger number of seismic stations around the volcano. It was a good idea but we had only four stations during the eruption; the eruption was over already, and it was impossible to install additional stations going “back to the past.” The data included in this book are heterogeneous enough but they are unique to describe the important volcanic events, and they give the understanding of the history of our monitoring of volcanic eruptions.

Eight years after the First Edition is not so long interval of time. But the number of publications in *Volcanic Seismology* during the first decade of the XXI century was comparable with the number of publications in this field during all time of the previous observations on volcanoes. So I understood that the new, updated and significantly expanded, edition of the book, is necessary. I added new chapters about the theoretical and experimental fundamentals of Volcanic Seismology; about long-period and very long-period seismic signals, observed at volcanoes; about microearthquake swarms associated with explosive and effusive activity at volcanoes; about acoustic signals associated with volcanic processes; and about failed eruptions within volcanic regions. I significantly rewrote and expanded the chapters about the



seismic signals associated with explosive events and with pyroclastic flows, rockfalls, and lahars.

I hope that this book will serve successfully for new generation of volcano seismologists and be a good friend to all interested in the nature of volcanic processes.

**Vyacheslav M. Zobin**

# Preface to the First Edition

I just returned from my first field seismological practice in Altai. It was October 1964, and I was a 20-year-old student of the Faculty of Geology and Geophysics at Novosibirsk State University in Siberia, Russia. Our guys told me that one man from Kamchatka wanted to see me. He was Dr. Pavel Tokarev, who was a chief of a new department for seismic monitoring of volcanic activity in the Institute of Volcanology and recruiting personnel for his department. He invited me to go to Kamchatka the following summer for my second field seismological practice at volcanoes and proposed the following theme for my year's essay: "Volcanic earthquakes and their relationship with volcanic eruptions."

I went to our university library and found a few papers concerning this problem. Among them were a famous paper by Takeshi Minakami (1960) about the prediction of volcanic eruptions, the papers by Georgy Gorshkov and by Pavel Tokarev about seismic study of Klyuchevskoy and Bezymianny volcanoes in Kamchatka, and so on. Thus I understood that this field of science was quite open for researchers, and I was lucky to enter to this "terra incognita."

I wrote this essay, presented it at the university student conference, received as a prize a beautiful book of *Elementary Seismology* by Charles Richter, and next summer went to my Kamchatka practice. I met a wonderful land. I climbed Avachinsky volcano, I had adventures at the flank cones of Klyuchevskoy volcano, and I decided to work there, in Kamchatka. Later, I wrote my MS thesis on the seismic wave attenuation beneath Klyuchevskoy volcano and was then accepted in the staff of the Institute of Volcanology in Petropavlovsk-Kamchatsky. Two months later, in October 1966, I had my first eruption of the Piip flank cone of Klyuchevskoy volcano; my first publication (together with P. Tokarev and V. Shirokov) about seismic signals associated with this eruption appeared in 1968. I began my life in volcanic seismology.

I spent in Kamchatka about 30 years. In 1971 the General Assembly of the International Union of Geophysics and Geodesy took place in Moscow. It was a great opportunity to meet many outstanding volcanologists of the world. Takeshi Minakami and Izumi Yokoyama became my friends; I was happy to receive Christmas cards from them for many years. In 1975 the great Tolbachik fissure eruption began in Kamchatka. A study of seismic

signals related to this eruption strongly improved our knowledge of the nature of volcanic earthquakes. I wrote a book about the source properties of volcanic earthquakes, and it was published in Russian in Moscow in 1979. It was one of the first books devoted to volcanic seismology.

I have lived in México for the last 9 years. I have prepared and delivered twice the courses in *volcanic seismology* for graduate students at the scientific center CICESE in Baja California, México and also have decided to write a book which you can have it now.

**Vyacheslav M. Zobin**

In the following photos, I am with my teachers and my students. One photo was taken in Moscow during the 1971 IUGG Assembly. The second in the first row (from the left) is Georgy Gorshkov, who was elected as a President of IAVCEI that year; the fourth is Takeshi Minakami. I am the fourth in the third row. The last in the fourth row is Izumi Yokoyama.



In the following other two photos, I am with the graduate students during my visits to Alaska Volcano Observatory, Fairbanks, United States in September 2010 and to King Saud University, Riyadh, Saudi Arabia in May 2011. It was a good opportunity to give them the revised version of my lectures on *volcanic seismology*.



# Acknowledgments

I thank my colleagues from Kamchatka, Russia and Colima, Mexico for their collaboration and useful discussions. My coauthors in the investigations in volcanic seismology Valentina Gorelchik, Pavel Tokarev, Valeria Levina, and Pavel Firstov helped me to understand better many aspects being discussed in this book. Seismic data obtained by the Kamchatkan regional seismic network and the Colima seismic network RESCO gave me an experience in the study of seismic signals recorded at volcanoes. The catalogs of volcanic earthquakes provided by Abdullah Al-Amri (Harrat Lunayyir, Saudi Arabia), Stefano Gresta and Domenico Patané (Etna volcano, Italy), Zenon Jiménez (El Chichon volcano, Mexico), Jun'ichi Miyamura (Usu volcano, Japan), Jennifer Nakata, David Shelly and Brian Shiro (Kilauea volcano, Hawaii), Kodo Umakoshi (Unzen volcano, Japan), Javier Lermo-Samaniego (Popocatepetl volcano, Mexico), and Willy Aspinall (Soufrière Hills volcano, West Indies); as well as the catalog of earthquakes of Pinatubo volcano (Philippine) prepared by the US Geological Survey team; the 1980 and 2004 catalogs of earthquakes of Mount St. Helens volcano (Cascades, United States) prepared by the Pacific Northwest Seismic Network, the Kamchatka Regional Earthquake Catalog; the catalog of earthquakes of Icelandic Meteorological Office (Grimsvøtn volcano); the catalog of Instituto Geográfico Nacional (El Hierro volcano); and the catalog of Harvard CMT solutions were very important in the preparation of this book. Galina Kopylova sent me the Catalog of Kamchatkan hot springs; Vladimir Leonov sent me the Catalog of Kamchatkan high-temperature geothermal systems. I use in this book the seismic records and video images obtained at Colima Volcano Observatory by the teams headed by Gabriel Reyes, Raúl Arambula, and Mauricio Bretón, respectively. Anastasia Chebrova, Luca D'Auria, Susanna Falsaperla, Pavel Firstov, Diego Gómez, Enrique Guevara, Bambang P. Istadi, Orlando Macedo, Stephan Malone, William Melson, Jun'ichi Miyamura, Yasuaki Sudo, and Mae Yamamoto provided me with the seismic records of volcanic events. Miguel González, Justo Orozco, and Armando Tellez helped me with the data acquisition. The processing of the digital signals was performed using the program DEGTRA provided by Mario Ordaz and SEISMO-VOLCANALYSIS provided by Philippe Lesage. Izumi Yokoyama, Daisuke Shimozuru, and Viktor Ivanov

sent me the biographical information and photos of the founders of volcanic seismology.

My work was benefited by numerous comments made during the preparation of the papers that formed the basis for this book. Stefano Gresta, Kazuhiro Ishihara, Sharon Kedar, Fred Klein, Jim Luhr, Steve McNutt, Seth Moran, Alejandro Nava, and Steven Sherburn, among others, helped me to better understand many aspects of volcanic seismology. Nick Varley read the text, prepared for the first edition, and made a gigantic work polishing my English. The comments of Andrea Cannata, Jonathan Fink, Hiroyuki Kumagai, Wendy McCausland, Robin Matoza, Silvio De Angelis, Aldo Martuzano, Natalia Stalbout, Nick Varley, and Gregory Waite helped me to improve the text.

I thank also all copyright holders, who allowed me to reproduce figures in this book: American Geophysical Union, Earthquake Research Institute of Tokyo University, Elsevier Science (Academic Press), Geological Society of London Publishing House, Institute of Volcanology and Seismology, Instituto de Geofísica, Istituto Nazionale di Geofisica, Seismological Society of America, Nature Publishing Group, Società Geologica Italiana, Springer-Verlag, Tohoku University, John Wiley and Sons, University College Dublin, US Geological Survey, Department of the Interior/USGS, and Volcanological Society of Japan. I thank S. De Angelis, M. Iguchi, V. Ivanov, M. Kasahara, O. Macedo, S. Matsumura, H. Okada, K. Tanaka, and H. Yamasato for the kind permission to reproduce their published figures.

Finally, I want to say that I have prepared this book considering that the basic volcanological and seismological ideas and theories may be taken from special textbooks. So I recommend to my readers some of the following textbooks in seismology and volcanology:

Lay, T., Wallace, T.C., 1995. *Modern Global Seismology*. Academic Press, San Diego, pp. 521.

Stein, S., Wysession, M., 2003. *An Introduction to Seismology, Earthquakes, and Earth Structure*, Blackwell Publishing, Oxford, pp. 498.

Francis, P., 1998, *Volcanoes. A Planetary Perspective*, Clarendon Press, Oxford, pp. 443.

Schmincke, H.-U., 2004. *Volcanism*. Springer, Berlin, pp. 324.

I also recommend the following review papers:

Chouet, B.A, 2003. Volcano seismology. *Pure and Applied Geophysics* 160, 739–788.

Kawakatsu, H., Yamamoto, M., 2015. Volcano seismology. *Treatise on Geophysics*, second ed. vol. 4, 389–419.

- Kumagai, H., 2009. Volcano seismic signals, source quantification of. In: Meyers, R.A. (Ed.), *Encycl. Complexity and System Sci.*, Springer-Verlag, Berlin, pp. 9899–9932.
- Matoza, R.S., Roman, D.C., 2022. One hundred years of advances in volcano seismology and acoustics. *Bulletin of Volcanology*, 84, 86.
- McNutt, S.R., Roman, D.C., 2015. Volcanic seismicity. In: Sigurdsson, H. (Ed.), *Encyclopedia of Volcanoes*, second ed. Academic Press, San Diego, pp. 1012–1033.
- Newhall, C. G., 2015. *Volcanology 101. Treatise on Geophysics*, second ed. vol. 4, pp. 355–387.
- Japanese readers can also use the book on volcanic seismology published in Japanese:
- Nishimura, T., Iguchi, M., 2006. *Volcanic Earthquakes and Tremor in Japan*, Academic Press, Kyoto University, Kyoto, pp. 242.

# Chapter 1

## Introduction

*The subject of **volcanic seismology** is an interesting, but the most difficult subject to study for seismologists. This is because seismic sources in volcanoes involve dynamic motion of gas, fluid and solid, and propagation paths in volcanoes are usually extremely heterogeneous, anisotropic, and absorptive, with irregular topographies and interfaces including cracks of all scales and orientations. Thus volcanic seismology is the most challenging to seismologists requiring ingenuity in designing experiments and interpreting observations.*

K. Aki, 1982. From the preface to the collection of papers “Volcanic Seismology.” IAVCEI Proceedings in Volcanology, vol. 3. Springer-Verlag, Berlin Heidelberg New York.

Volcanic seismology is a science about seismic signals originating from volcanoes and associated with volcanic activity. The study of the origin of these signals, their spatial–temporal distributions, their relationships with volcanic processes, and using them as an instrument to investigate the volcano deep structure and to predict a volcanic eruption together comprise the subject of volcanic seismology.

### 1.1 TERMS AND DEFINITIONS

The subject of volcanic seismology is about the interaction between volcanic and seismotectonic processes. Therefore, it is necessary to define the main terms of volcanology, tectonics, and seismology at the very beginning of the book.

#### 1.1.1 Volcanic Terms

A volcano is a vent or chimney, which connects a reservoir of molten matter known as magma, in the depths of the crust of the Earth, with the surface of the Earth. The material ejected through the vent frequently accumulates around the opening, building up a volcanic edifice. The material ejected consists of liquid lava and broken fragments of partially or completely solidified rock (pyroclastic debris), as well as great quantities of gases. The gases are the motivating force and the most important factor in volcanic action (Bullard, 1963).



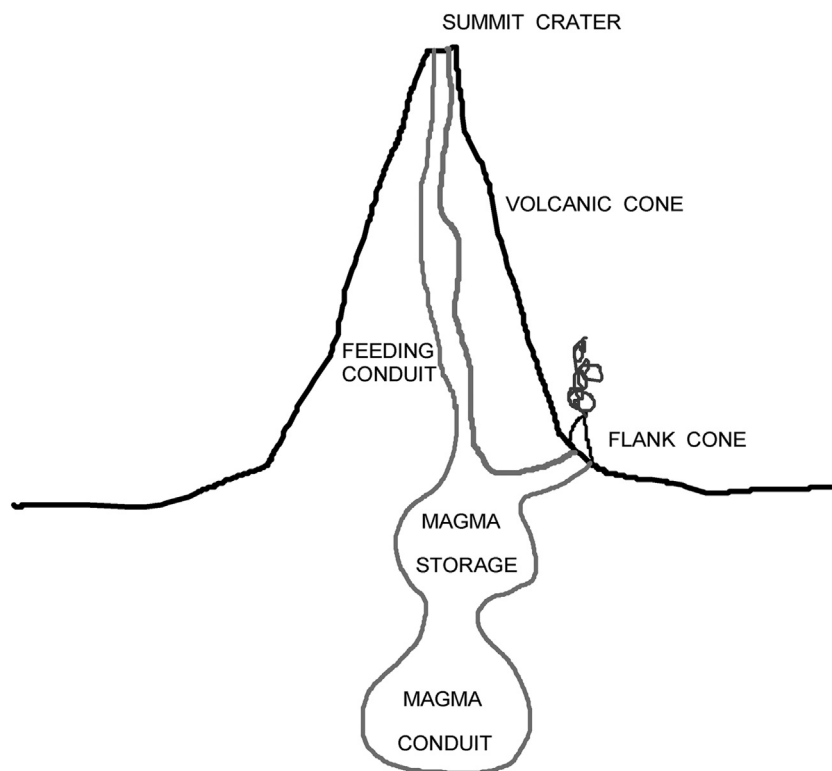


**FIGURE 1.1** The main elements of volcanic edifice.

Fig. 1.1 shows the main elements of a volcano. A volcanic cone is the result of the accumulation of the ejected material around the vent. A crater is the surface connection of the volcanic conduit through which the ejected material reaches the surface. The crater may be situated on the summit of the cone (summit crater) or on a slope of the cone (flank crater). An enormous crater representing a large depression in the truncated summit of volcano is called caldera.

Beneath a volcano, there is a magma feeding system of the volcano that consists of (1) a deep magma reservoir, or a chamber, which is a reservoir of magma in the shallow part of lithosphere, from which volcanic material is derived; (2) an intermediate magma storage, a place between the magma chamber and the Earth's surface, where magma may be collected before an eruption; and (3) a feeding conduit, connecting the magma collectors with the surface. The surface appearance of magma is called lava (Bullard, 1963). Volcanoes are widely distributed on the surface of the Earth and on the ocean floor (Fig. 1.2). Volcanic belts go along subduction zones as well as along oceanic ridges. Simkin and Siebert (1994) have noted more than 1300 active volcanoes with more than 5500 dated eruptions.



**FIGURE 1.2** Major lithospheric plates and global distribution of some active and dormant volcanoes. *Courtesy: US Geological Survey.*

A volcanic eruption is a process of material transport from the Earth's depths to the surface. [Figs 1.3–1.5](#) show the typical surface manifestation of volcanic features during the activity of basaltic, andesitic, and dacitic volcanoes. According to their origin within the volcanic edifice, the volcanic eruptions may be of (1) central (or summit) type with an eruption of the summit crater; (2) flank (or lateral) type with an eruption of the flank crater; or (3) fissure type, when the eruption takes place from an elongated fissure, rather than from a central vent.

The magnitude of eruption is described by Volcanic Explosivity Index (VEI) that combines the total volume of explosive products, the eruptive cloud height, descriptive terms, eruption type, duration, and so on and has the grades ranging from 0 to 8 ([Newhall and Self, 1982](#)). [Newhall and Self \(1982\)](#) proposed the range of VEI for five main types of volcanic eruptions: Hawaiian (VEI 0–1), Strombolian (VEI 1–2), Vulcanian (VEI 2–4), Plinian (VEI 4–8), and Ultra-Plinian (VEI 5–8). The first two types of eruptions, Hawaiian and Strombolian, are common for basaltic volcanoes. Vulcanian eruptions may be observed mainly at andesitic and dacitic volcanoes. Plinian and Ultra-Plinian eruptions occur at dacitic and rhyolitic volcanoes. During an eruptive process, the type and style of activity and the nature and type of products can change ([Table 1.1](#)) ([Newhall and Self, 1982](#); [Vergnolle and Mangan, 2000](#)).

### 1.1.2 Tectonic Terms

The movement of magma in the interior below the volcanic structure leads to the change in hydrostatic equilibrium and to the fracturing of the Earth material. Tectonic faults and fractures represent the breaks in rock material of

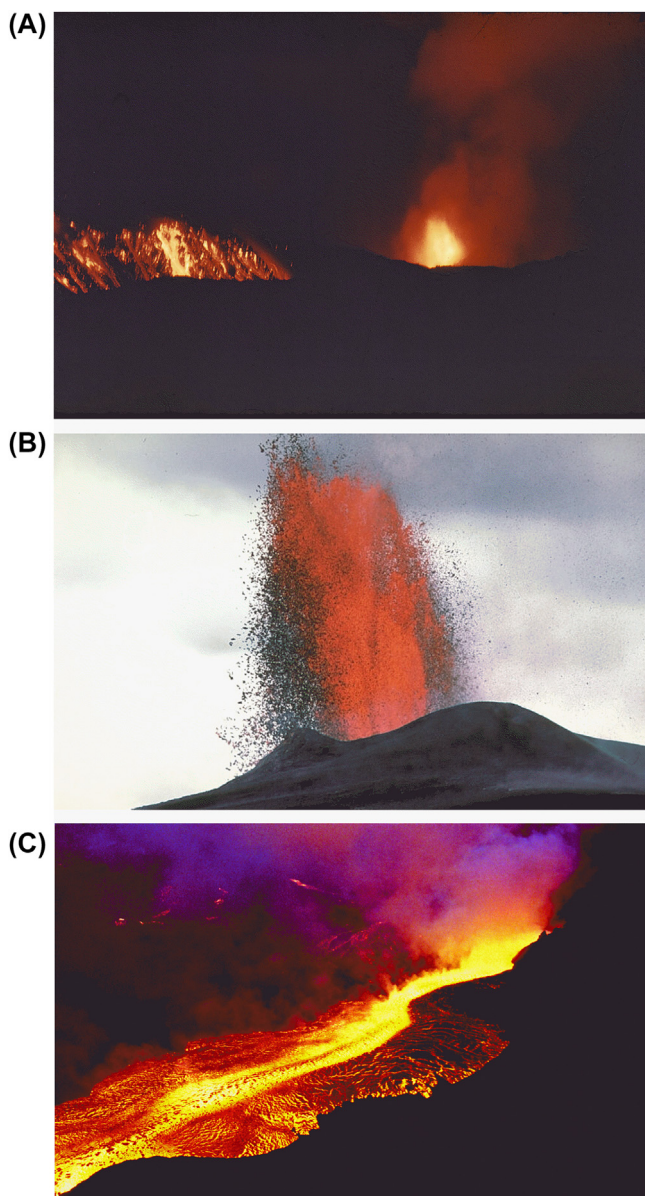


**FIGURE 1.3** Surface manifestation of volcanic features during the activity of basaltic volcanoes. (A) The 1975 fissure eruption of New Tolbachik volcanoes, Kamchatka. Eruption from the new cone and its lava flow. (B) Lava fountain and spatter cone on Kilauea Volcano, Hawaii. (C) The November 1989 lava flow on Etna volcano. (A) Photograph by V. Zobin. (B) Photograph by J.D. Griggs; Courtesy: U.S.G.S. (C) Photograph by A. Amantia; Courtesy: INGV Catania.

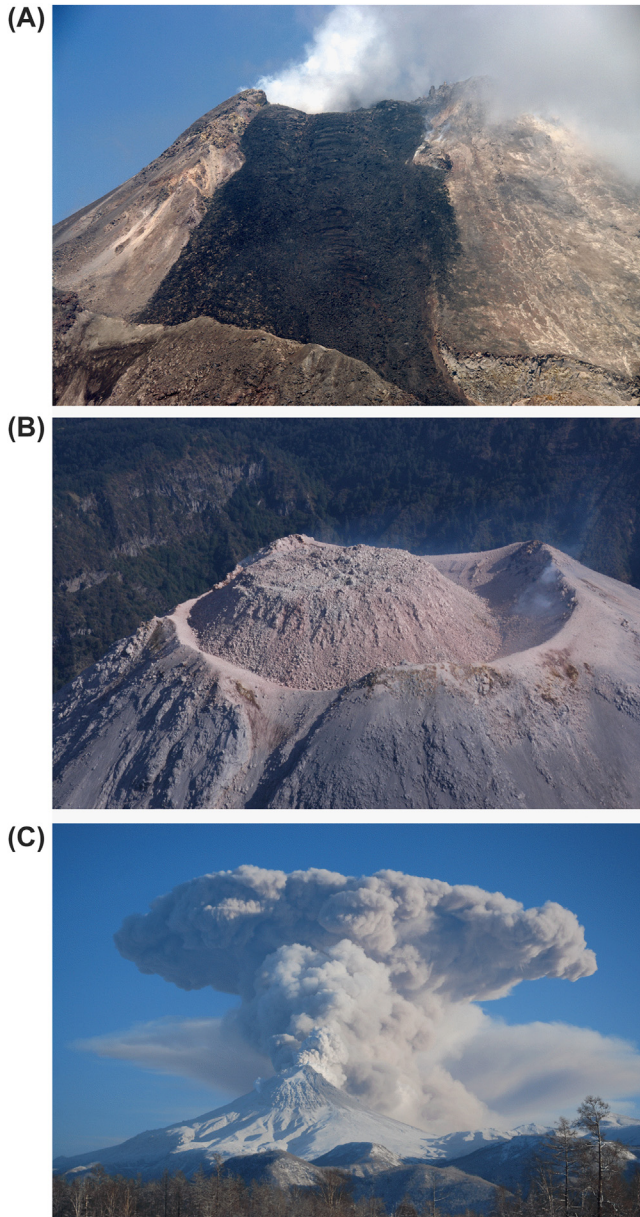


**FIGURE 1.4** Surface manifestation of volcanic features during the activity of andesitic volcanoes. (A) Block lava flow moving down the flank. Bezymianny volcano, Kamchatka. (B) Lava dome in the crater of Volcán de Colima. (C) Explosion at Kizimen volcano, Kamchatka. (A) Photograph by P. Izbekov; Courtesy: Alaska Volcano Observatory. (B) Photograph by M. Bretón; Courtesy: Colima Volcano Observatory. (C) Photograph by E. Vlasov; Courtesy: the Institute of Volcanology and Seismology.



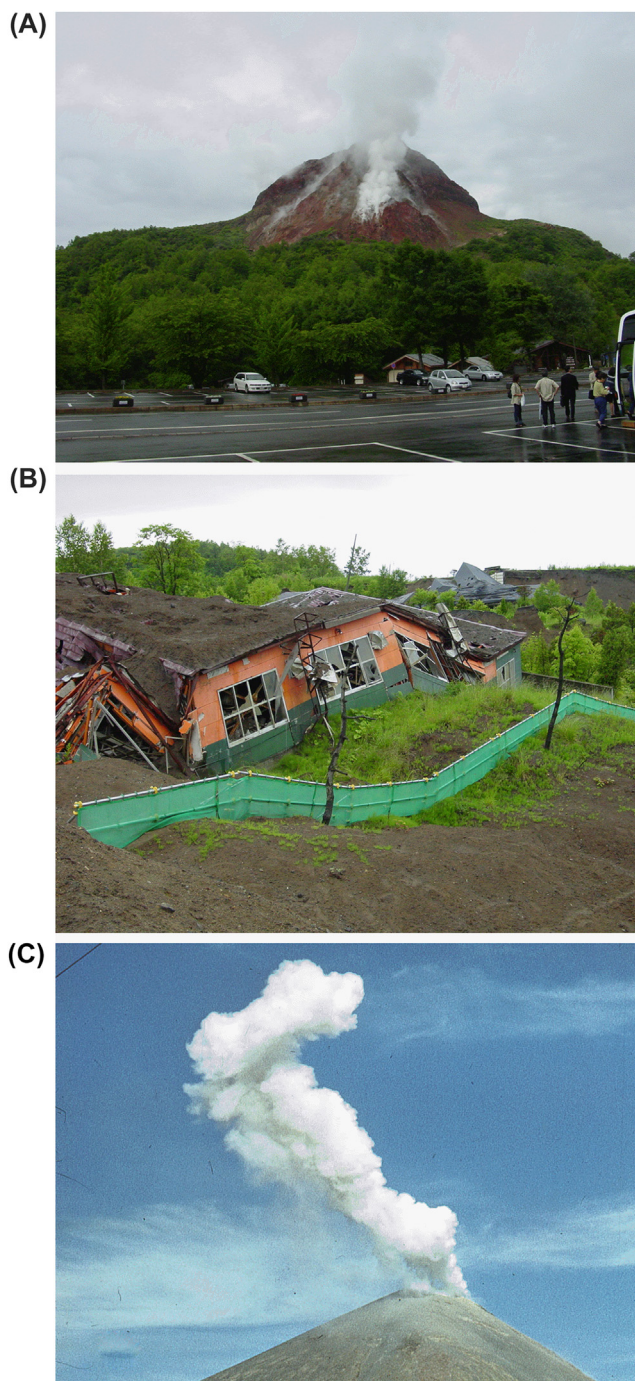


**FIGURE 1.5** Surface manifestation of volcanic features during the activity of dacitic volcanoes. (A) Lateral lava dome Showa Shinzan (Usu volcano, Hokkaido). (B) Surface deformation caused by the subsurface moving of dacitic lava. The 2000 lateral eruption of Usu volcano, Hokkaido. (C) Explosive activity at Karymsky volcano, Kamchatka. *Photographs by V. Zobin.*

**TABLE 1.1** General Characteristics of the Main Types of Volcanic Eruptions (Newhall and Self, 1982; Vergnolle and Mangan, 2000).

| Type of Eruption | Volume of Ejecta (m <sup>3</sup> ) | Column Height (km) | Type of Magma                        | Eruptive Style          | Volcanic Features                      |
|------------------|------------------------------------|--------------------|--------------------------------------|-------------------------|----------------------------------------|
| Hawaiian         | 10 <sup>4</sup> –10 <sup>6</sup>   | 0.1–1              | Basaltic                             | Fissure, central, flank | Lava flow, fire fountain               |
| Strombolian      | 10 <sup>4</sup> –10 <sup>7</sup>   | 0.1–5              | Basaltic                             | Central                 | Lava flow                              |
| Vulcanian        | 10 <sup>6</sup> –10 <sup>9</sup>   | 1–25               | Andesitic, andesitobasaltic, dacitic | Central, flank          | Lava flow, pyroclastic flow, lava dome |
| Plinian          | 10 <sup>8</sup> –>10 <sup>12</sup> | 10–>25             | Rhyolitic, trachytic, phonolitic     | Central                 | Lava dome, pyroclastic flow            |
| Ultra-Plinian    | 10 <sup>9</sup> –>10 <sup>12</sup> | >25                | Rhyolitic, trachytic                 | Central                 |                                        |

*Source:* From Newhall, C.G., Self, S., 1982. The Volcanic Explosivity Index (VEI): an estimate of explosive magnitude for historical volcanism. J. Geophys. Res. (Oceans and Atmospheres) 87, 1231–1238.

different scales. We define fractures as the ruptures of length less than 100 m; greater ruptures would be faults. As a rule, the system of fractures forms around general faults and reflects the local rupturing of different parts of the fault. A type of movement along the fault plane characterizes tectonic faults. There are three main types of tectonic movements: normal faulting, thrust, and strike-slip. The normal fault and the thrust are characterized by a low angle of inclination with reference to a horizontal plane. With normal faulting, the hanging wall of the fault has been depressed, relative to the footwall; the thrust is a reverse fault. The strike-slip represents the horizontal movement along the vertical plane. Volcanoes usually lie along a tectonic fault or intersection of faults, usually within a rift structure where normal faulting is dominant.

### 1.1.3 Seismological Terms

The rupturing along the fractures and faults as well as the eruptive activity produces the seismic waves, which could vibrate the Earth surface as earthquakes. The site of initial radiation of seismic waves in the interior is earthquake focus, or hypocenter; its projection to the surface is epicenter.

The seismic waves are divided into body waves and surface waves. The body waves  $P$  (compressional waves with volumetric changes) and  $S$  (shear waves without volume change) radiate from the earthquake focus, while the surface Rayleigh and Love waves are formed as a result of the interaction of body waves with the surface of the Earth. The record of seismic waves is a seismogram (Fig. 1.6). The earthquake size is described by magnitude  $M$  (or  $m$ ), the conventional value proportional to the earthquake energy, and by seismic moment  $M_0$ . The focal mechanism characterizes the stress system acting in the earthquake source zone, while the stress drop gives the value of stress release during the earthquake.

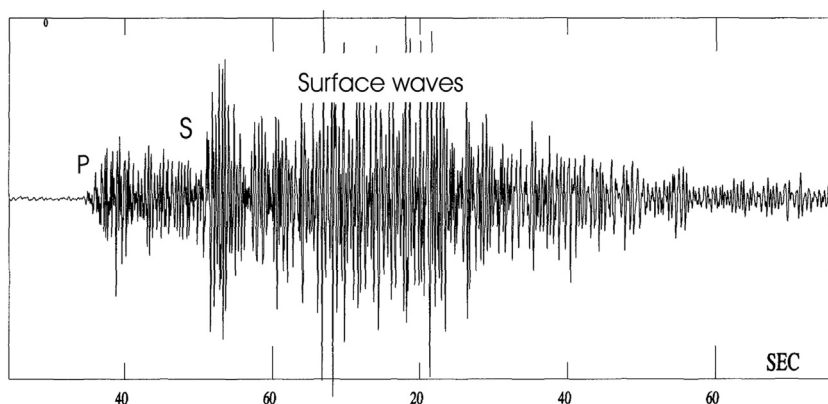


**FIGURE 1.6** Seismogram of the shallow local earthquake. Vertical component of a short-period instrument. *Courtesy: Colima Volcano Observatory.*

## 1.2 SUBJECT OF THE BOOK

Volcanic seismology does not yet have any plain theory; it is more a science of observations and cumulative experience. The quantitative theory of seismovolcanic processes is being developed today only for some cases, and it is far from complete. This book is an introduction to *Volcanic Seismology*. It is devoted to the study of the relationships between volcanic earthquakes and eruptive process and of the nature of volcanic earthquakes.

The book consists of 21 chapters. Chapters 2 and 3 introduce us to the history and the topic of volcanic seismology and discuss the theoretical and the experimental models that were developed for the study of the origin of volcanic earthquakes.

Chapters 4–7 give the theoretical basis for the occurrence of volcano-tectonic earthquakes and their swarms and describe the case stories of volcano-tectonic activity associated with the eruptions at basaltic, andesitic, and dacitic volcanoes. For this book, the author selected 40 cases of volcanic eruptions (or episodes) at 20 volcanoes that occurred all over the world from 1910 to 2022. This selection was argued by different reasons. The most important of them was the possibility to have a catalog of seismic events for special analysis, the quality of seismic observations, and the historical significance of the eruption, among others. Chapters 8–10 describe the general regularities of volcano-tectonic earthquake swarms, their participation in the eruptive process, their source properties, and the hazard of strong volcano-tectonic earthquakes.

Chapter 11–18 are devoted to the description of eruption earthquakes. The author gives the theoretical basis for the occurrence of eruption earthquakes and describes the volcanic tremor, seismic signals associated with lava domes, pyroclastic and lava flows, rockfalls and lahars, explosion earthquakes, long-period and very long-period seismic signals at volcanoes, microearthquake swarms, and acoustic events. The author also discusses their relationship to the eruptive process.

Chapters 19 and 20 discuss the mitigation of volcanic hazards and some specific appearances of seismicity in volcanic regions. The author writes briefly about the methodology of seismic monitoring of volcanic activity and the experience in forecasting of volcanic eruptions by seismic methods. Chapter 21 talks also about the seismic activity in the regions of dormant volcanoes and presents the review of natural seismicity associated with the hydrothermal activity within the volcanic environment.

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## 10 Volcanic Seismology

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## Chapter 2

# Seismicity at Volcanoes

This Chapter presents the history of seismic monitoring of volcanic activity. The classification of volcanic earthquakes is given. The types of seismic sequences associated with volcanic activity are illustrated.

### 2.1 HISTORY OF SEISMIC MONITORING OF VOLCANIC ACTIVITY

Earthquakes were associated with volcanic eruptions from ancient times. Pliny the Younger gave us the first scientific description of the eruption of Vesuvius in AD 79. He wrote about numerous earthquakes related to this eruption. For many years, people considered volcanoes as the main origin of earthquakes in general. In the classification of earthquakes that was prepared in the beginning of the 20th century, earthquakes of volcanic origin were in the first place.

The Vesuvius volcano occupies a special place in the history of the study of seismic signals associated with volcanic activity. It was the first volcano whose earthquakes were mentioned in the scientific literature, it was the first volcano to have a volcano observatory in 1847, and it was the first to be monitored using seismological equipment. The Palmieri electromagnetic seismograph, built in 1862, was the first seismic instrument to record volcanic seismicity. The seismograms were written on a telegraph tape. Later, in 1903, the Bosch—Omori seismograph was installed at Mount Pelée, Martinique, 1 year after the great eruption of 1902 (Ferrucci, 1995). Seismic observations of volcano at Kilauea, Hawaii, started in 1912 by H. Wood just after the foundation of the Hawaiian Volcano Observatory (Wood, 1913, 1915; Macdonald, 1972).

As a science, *volcanic seismology* was born when the Japanese seismologist Fusakichi Omori (Box 2.1) began his investigations of seismic signals related to the 1910 eruptions of Usu and Asama volcanoes (Omori, 1912, 1911–1913) and the 1914 eruption of Sakurajima volcano (Omori, 1914, 1922). Omori (1912) defined the volcanic earthquake as “seismic disturbance, which is due to the direct action of the volcanic force, or one whose origin lies under, or in the immediate vicinity of, a volcano, whether active, dormant, or extinct.”

The three-component seismic station, with a magnification of  $\times 100$  and the period of the pendulum about 4 s, was installed by Omori in the period

## BOX 2.1 Omori



Prof. Fusakichi Omori (1868–1923) was a founder of volcanic seismology. He began his work in seismology with John Milne and became in 1896 the Professor of Imperial (Tokyo) University and Secretary of the Imperial Earthquake Investigation Committee. For decades, he was the natural leader in all Japanese investigations on volcanoes and earthquakes. Omori's instrumental observations of seismic events during the 1910 Usu, the 1910–14 Asama, and the 1914 Sakurajima eruptions showed the great significance of seismic monitoring of volcanic eruptions for the understanding of volcanic processes and forecasting eruptions. Prof. Omori left us the detailed descriptions of these seismic events; researchers still use his papers in volcanic seismology. His death was dramatic. D. Shimozuru wrote to me in 1999: "It is well known the dispute between Prof. Omori and Prof. Imamura about the possibility of an occurrence of a great earthquake in the Tokyo area in 1923. During his participation in the Congress at Australia, the great Kanto earthquake occurred on September 1, 1923. Prof. Omori was defeated. He was heartbroken during his return voyage from Australia, and soon after he passed away."

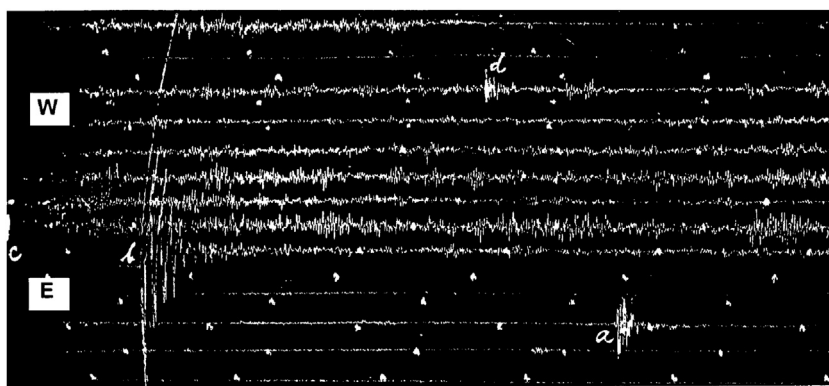
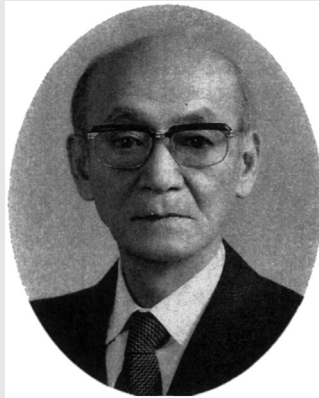


**FIGURE 2.1** Seismogram of volcanic earthquakes (marked as *a*, *b*, and *d*) and volcanic tremor (continuous signal) recorded during August 7, 1910, to August 8, 1910, at the Sobetsu seismic station near Usu volcano, Hokkaido, Japan, with a magnification of  $\times 100$ ; component WE. From Omori, F., 1911–1913. *The Usu-san eruption and earthquake and elevation phenomena. Bulletin of the Imperial Earthquake Investigation Committee* 5, 1–38.

from July 1910 to August 1910 at a distance of about 1–2 km from the craterlets of Usu. This station allowed him to record the earthquakes originating from the volcano and the volcanic microtremors whose period was about 0.5 s (Fig. 2.1). The microtremors were observed only during volcanic activity. Omori demonstrated a good relation between the appearance of microtremors and explosions of the volcano.

In the same paper, Omori (1911–1913) noted also that many eruptions of Japanese volcanoes were preceded by numerous earthquakes. In such a case, he wrote, “seismograph observations made at the vicinity of the center of volcanic activity would give people a warning of the approaching outburst.” Analysis of volcanic earthquakes, recorded during volcanic activity by Omori, had shown the great potential of seismic observations. They became the main instrument for volcanic activity monitoring. Among the most important stages in the development of *volcanic seismology*, we noted the foundation of seismic observations at Asama Volcano Observatory on

### BOX 2.2 Minakami



Prof. Takeshi Minakami (1909–85) graduated from Tokyo University and at the age of 25 became the first director of the first Japanese Volcano Observatory at Asama volcano. He initiated the seismological observations at Asama, and his research works of Asama volcanic earthquakes allowed him to elaborate their classification. He developed a methodology for the statistical forecasting of volcanic eruptions at Asama by seismic data. His 1960-year paper “Fundamental research for predicting volcanic eruptions” helped to turn a volcanic seismology into a science. Later, T. Minakami studied seismic events associated with the eruptive activity of Usu, Miyake-sima, and Arenal volcanoes. As a Professor of Earthquake Research Institute of Tokyo University, he prepared a new generation of Japanese volcano-seismologists. D. Shimozuru wrote in 1971 that Prof. Minakami “obtained pioneering results, which led to great progress toward the better understanding of volcanic phenomena.”

Honshu Island, Japan, in 1934, and at Kamchatka Volcano Observatory in Klyuchi, Kamchatka, Soviet Union, in 1946.

Asama Volcano Observatory was established by Takeshi Minakami (Box 2.2) in 1934, and from the very beginning, it was equipped with continuously operated seismographs. The observations of volcanic and seismic activity at Asama volcano allowed Minakami to propose a classification of volcanic earthquakes into four types, which is the basis for the study of volcanic earthquakes even in the present day. He obtained a good correlation between the long-term variation in the number of volcanic earthquakes

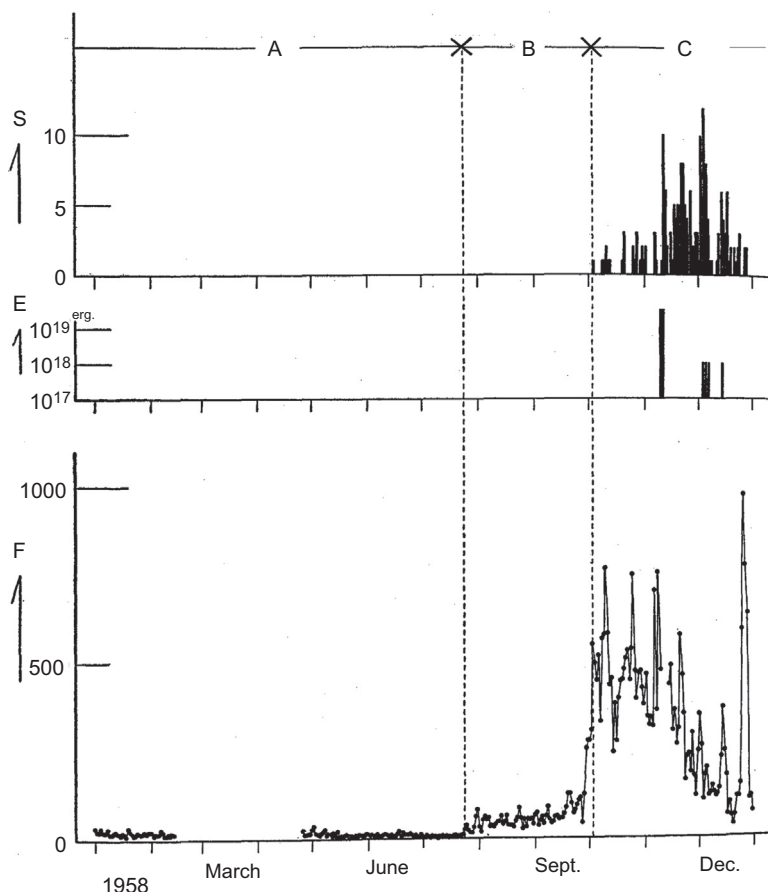
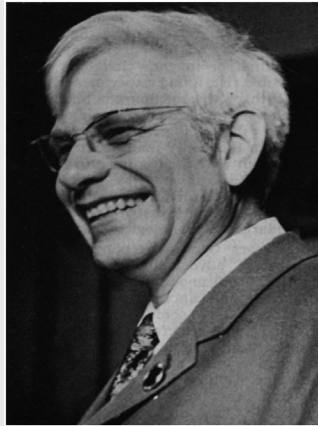


**FIGURE 2.2** Relation between eruptions and shallow earthquakes (B-type) of Asama, Honshu, Japan. A, B, and C indicate the calm, preeruptive, and explosive eruption stages of the volcano. *S* is the daily frequency of explosions, *E* is the kinetic energy of strong explosions, and *F* is the daily frequency of shallow earthquakes. From Minakami, T., 1960. *Fundamental research for predicting volcanic eruptions. Part I. Bulletin of the Earthquake Research Institute, Tokyo University* 38, 497–544.

and volcanic eruptions (Fig. 2.2). The increase of volcanic B-type earthquakes before the volcanic outburst became the criteria for the development of the method of eruption forecasting (Minakami, 1960).

The continuous seismological observations at Kamchatka Volcano Observatory in Klyuchi began in 1946. They were headed by Georgy Gorshkov and later by Pavel Tokarev. Gorshkov (Box 2.3) was the first to discover that *S*-waves of distant earthquakes greatly attenuate the volcano; it allowed him to estimate the depth of the upper mantle magma chamber of Klyuchevskoy volcano (Gorshkov, 1956). The great eruption of Bezymianny volcano in 1955–56 gave numerous data for the detailed analysis of the

### BOX 2.3 Gorshkov



Dr. Georgy Stepanovich Gorshkov (1921–75) entered volcanic seismology in the very beginning of his volcanological career. In 1948, he was appointed a Director of Kamchatka Volcano Observatory in the Klyuchi village. A seismic station was installed here in the end of 1946, and Gorshkov became the first who studied the seismic records of the Kamchatka volcanoes. Two of his seismological results were fundamental. He discovered the screening effect of magma on *S*-waves of distant earthquakes and determined the depth of the magma chamber of Klyuchevskoy volcano in the upper mantle. He studied in detail the relationships between seismic energy release and volcanic activity associated with the great 1955–56 eruption of Bezymianny volcano. Dr. Gorshkov worked in volcanic seismology only for about 10 years. He was a petrologist, and later, he worked mainly in this area of volcanology. His monographs “Bezymianny Volcano and the Peculiarities of its Last Eruption, 1955–1963” (1965; with G.E. Bogoyavlenskaya) and “Volcanism and Upper Mantle” (1969) became the important events in a volcanological science. At the end of his life, Georgy Gorshkov was elected as the President of the International Association of Volcanology and Chemistry of the Earth’s Interior.

interrelation of seismic and volcanic events. This was the first complete history of a great eruption described in terms of seismic information (Gorshkov and Bogoyavlenskaya, 1965). The following analysis of the seismic activity related to the annual small eruptions of Bezymianny in 1957–61 allowed Tokarev (Box 2.4) to develop a method of forecasting of volcanic eruptions based on a change in energy release of the preceding seismic events (Fig. 2.3).

A network of high-gain seismometers at Hawaii Volcano Observatory has been operated on Kilauea volcano since the 1950s, and up to 1985, it consisted of 50 seismic stations (Klein et al., 1987). The seismic studies allowed for the construction of a fine model of the Earth's crust near the Hawaiian volcanoes. The basic model of Kilauea was proposed (Eaton and Murata,

#### BOX 2.4 Tokarev



Dr. Pavel Ivanovich Tokarev (1923–93) graduated from Moscow University, Russia, and began his scientific activity in the Laboratory of Volcanology in Moscow in the middle of the 50s. His first papers were published in 1958–59 and were devoted to the study of the structure of the Kurile-Kamchatka subduction zone and the relationship between seismicity and volcanism in this zone. In 1958, he moved to Kamchatka and began to work at the Kamchatka Volcano Observatory in the village of Klyuchi. Tokarev installed the first seismic network for the monitoring of the Klyuchevskaya group of volcanoes and began to develop the method of seismic forecasting of volcanic eruptions for Bezymianny volcano. His first publications in this field appeared in 1963–71 and became a classic. Later, Tokarev headed the Department of Geophysical Monitoring of Volcanic Activity in the Institute of Volcanology in Petropavlovsk-Kamchatsky and studied the seismic activity of Klyuchevskoy, Bezymianny, Karymsky, and New Tolbachik volcanoes. His forecasting of the 1975 fissure eruption of New Tolbachik volcanoes was published in a regional newspaper a few days before the beginning of the eruption and allowed a group of volcanologists to travel to the site of forthcoming eruption 1 week before the event begun. Pavel Ivanovich founded a system of seismic monitoring for volcanic activity in Kamchatka and prepared a team of seismo-volcanologists who continue now his efforts in seismic studies of volcanoes.

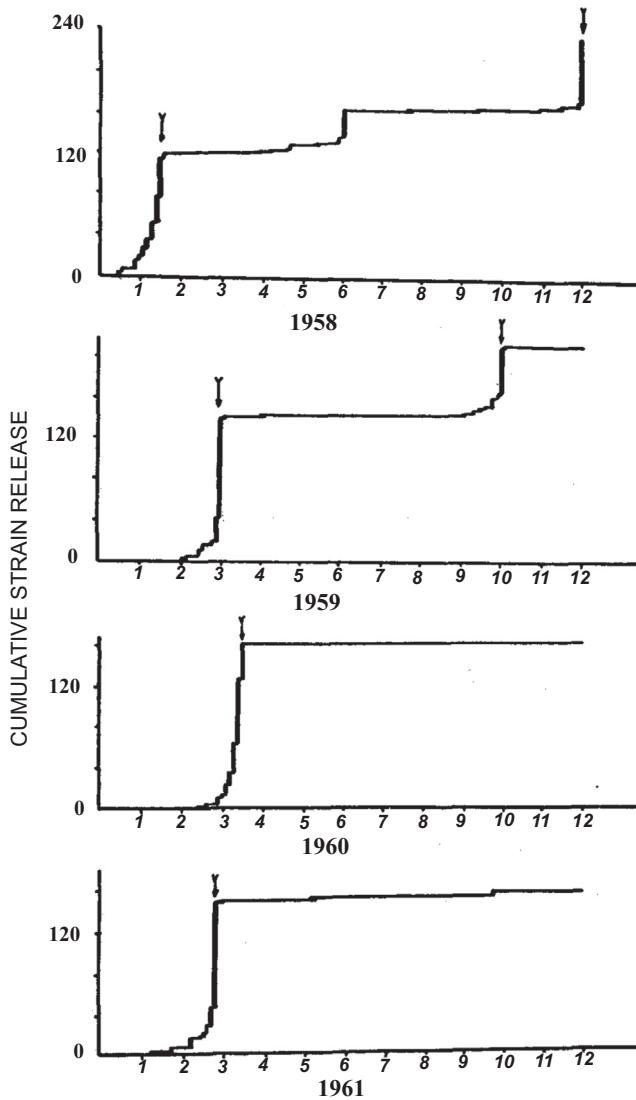


**FIGURE 2.3** Variations of the cumulative strain release [in  $\text{erg}/2$ ] recorded at Bezmianny volcano, Kamchatka in 1958–61. Arrows show the eruptions. The curves indicate an exponential increase in seismic energy release before an eruption. From Tokarev, P.I., 1963. *On a possibility of forecasting of Bezmianny volcano eruptions according to seismic data*. *Bulletin Volcanologique* 26, 379–386.

1960) with a magma reservoir at the depth of about 3–4 km beneath the caldera. The generalized histories of Kilauea earthquakes represent the detailed history of volcanic activity of Kilauea from 1970 (Klein et al., 1987). The regularities of the inflation–deflation cycle of the magma reservoir and the



appearance of a characteristic related to seismic events combined with ground deformation monitoring allowed for the development of the short-term forecasts of Hawaiian eruptions (Klein, 1984).

## 2.2 CLASSIFICATION OF VOLCANIC EARTHQUAKES

The first observations of volcanic earthquakes showed that the seismic signals originating from volcanic activity had different waveforms. Studying the seismic signals during the 1910 Usu eruption, Omori (1911–1913) wrote about “occurrence besides the proper volcanic earthquakes, of well-defined small quick unfelt vibrations, which may be termed micro-tremors” (see Fig. 2.1). The first and still very popular classification of volcanic earthquakes was proposed by Minakami (1960, 1974a). He divided them into four types (Fig. 2.4) according to the location of their foci, their relationship to the eruptions, and the nature of the earthquake motion.

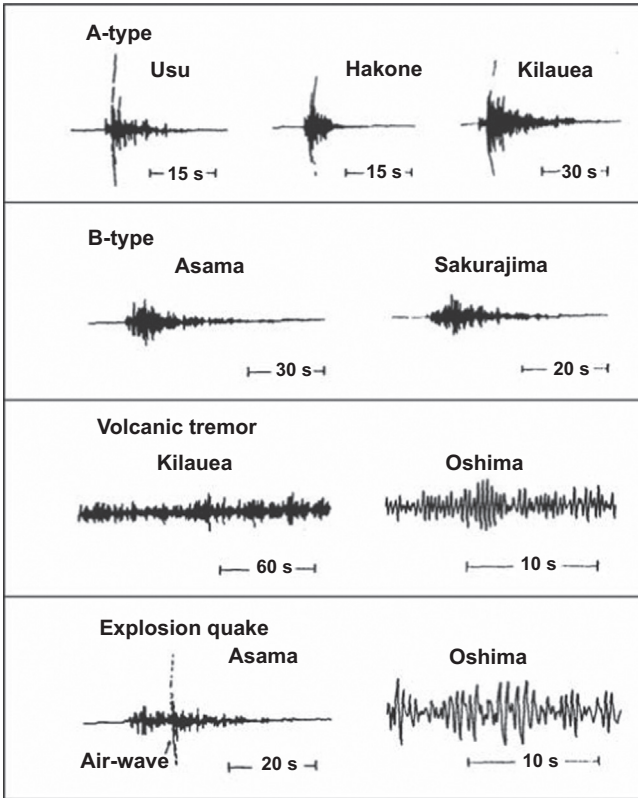


**FIGURE 2.4** An illustration of Minakami’s classification of volcanic earthquakes. From Minakami, T. (1974a). *Seismology of volcanoes in Japan*. In: Civetta, L., Gasparini, P., Luongo, G., Rapolla, A., (Eds.), *Physical Volcanology*, Elsevier, Amsterdam, pp. 1–27.

### 2.2.1 A-Type

These are earthquakes originated from the bases of volcanoes or from the depths of about 1–20 km. These earthquakes take place before and during the first stage of eruptive activity and occur in swarms. They are generally less than 6 in magnitude. The nature of earthquake motions cannot be distinguished from those of general shallow earthquakes (tectonic earthquakes). The *P*- and *S*-phases of seismic waves are also clearly defined.

### 2.2.2 B-Type

The hypocenters are limited to an area of about 1 km in radius around the active crater. The hypocenters of these earthquakes are shallower than those of the A-type volcanic earthquakes. These earthquakes take place in swarms in the shallow part ranging from Earth's surface to the depths of several hundred meters. Their magnitudes are generally extremely small. The surface waves are predominating, and the *S*-phase is not clear.

### 2.2.3 Explosion Earthquakes

These earthquakes accompany individual explosive eruptions. The amplitude of these earthquakes is related to the magnitude of the explosive eruptions. The earthquake motions show an exceeding predominance of longer wavelength as comparable with those of the A-type volcanic and tectonic earthquakes; the initial motion is “push” in every direction. On the seismograms of explosion earthquakes, we often find disturbances caused by air shocks. The hypocenters are situated beneath the active crater floor.

### 2.2.4 Volcanic Pulsation or Continuous Volcanic Microtremors

The main part of these vibrations consists of the surface waves. Volcanic tremor has a form of an irregular sinusoid of rather long duration compared with earthquakes of the same amplitude.

Volcanic earthquakes of all the four types can participate in eruptive activity during the same eruption. Fig. 2.5 illustrates the seismic history of the 1966 flank eruption of Klyuchevskoy volcano, Kamchatka (Tokarev et al., 1968). Prior to the eruption, volcanic earthquakes of A- and B-types were observed (bars 5). An eruptive fissure opened after the cessation of the earthquake sequence. In the beginning of the eruption, volcanic tremor (curve 3) started simultaneously with explosion earthquakes (curve 2). Volcanic tremor stopped just before the second swarm of B-type earthquakes (curve 4), but explosion earthquakes continued up to the end of the eruption.

Minakami's classification stays basic and popular up to today. Of course, for every volcano, seismologists use their own classification with a more

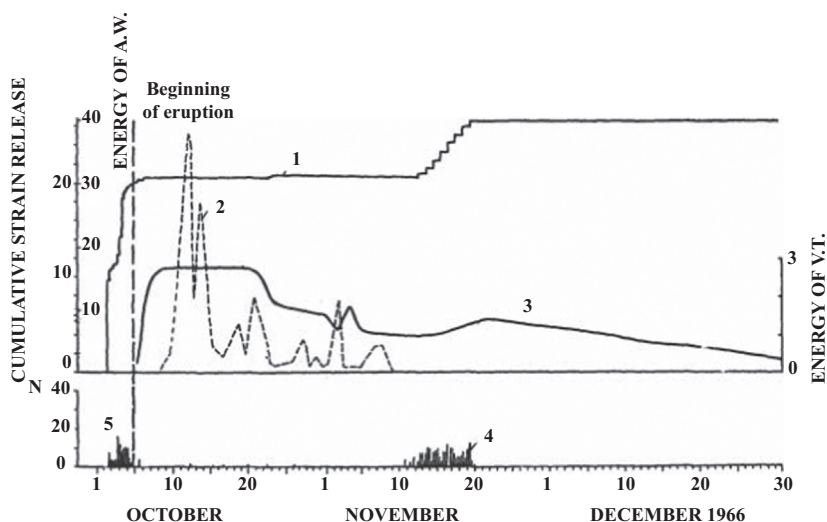


**FIGURE 2.5** Seismic history of the 1966 flank eruption of Klyuchevskoy volcano, Kamchatka. Curves 1–5 show the variations: (1) in the cumulative strain release measured in [10–8] erg/2 for A- and B-type earthquakes; (2) in the energy of air waves (AW) produced by explosions and measured in [10–15] erg/day; (3) in the energy of volcanic tremor (VT) measured in [10–14] erg/day; (4) in the number of B-type earthquakes of the second swarm; and (5) in the number of A- and B-type earthquakes of magnitude  $M > 0$ .  $N$  is the number of earthquakes. From Tokarev, P.I., Shirokov, V.A., Zobin, V.M., 1968. *Seismic events associated with the Piip lateral eruption in October–December, 1966. Bulletin of Volcanological Station 44, 30–41 (in Russian).*

detailed description of every subtype of earthquakes (Tokarev, 1966; Latter, 1979; Malone, 1983; Lahr et al., 1994; Harlow et al., 1996; Ratdomopurbo and Poupinet, 2000; McNutt, 1996, 2000).

A majority of the seismic signals defined by these authors were observed during the 1997–2017 unrest at andesitic Volcán de Colima, Mexico. Fig. 2.6 shows broadband seismic records of these volcanic earthquakes. Video monitoring allowed us to identify the nature of a majority of them. The seismic records of volcano-tectonic earthquakes were characterized by clear arrivals of  $P$ - and  $S$ -waves. The next two types of seismic records (EXP and HYB) were associated with Vulcanian explosive activity. EXP events with a low-frequency initial pulse were associated with dark magmatic explosions, while HYB events with a high-frequency initial pulse were mainly associated with white cloud degassing. Rare long-period signals were associated with small explosive events or were observed without any direct relation to eruptive activity. The swarms of short-duration and low-amplitude

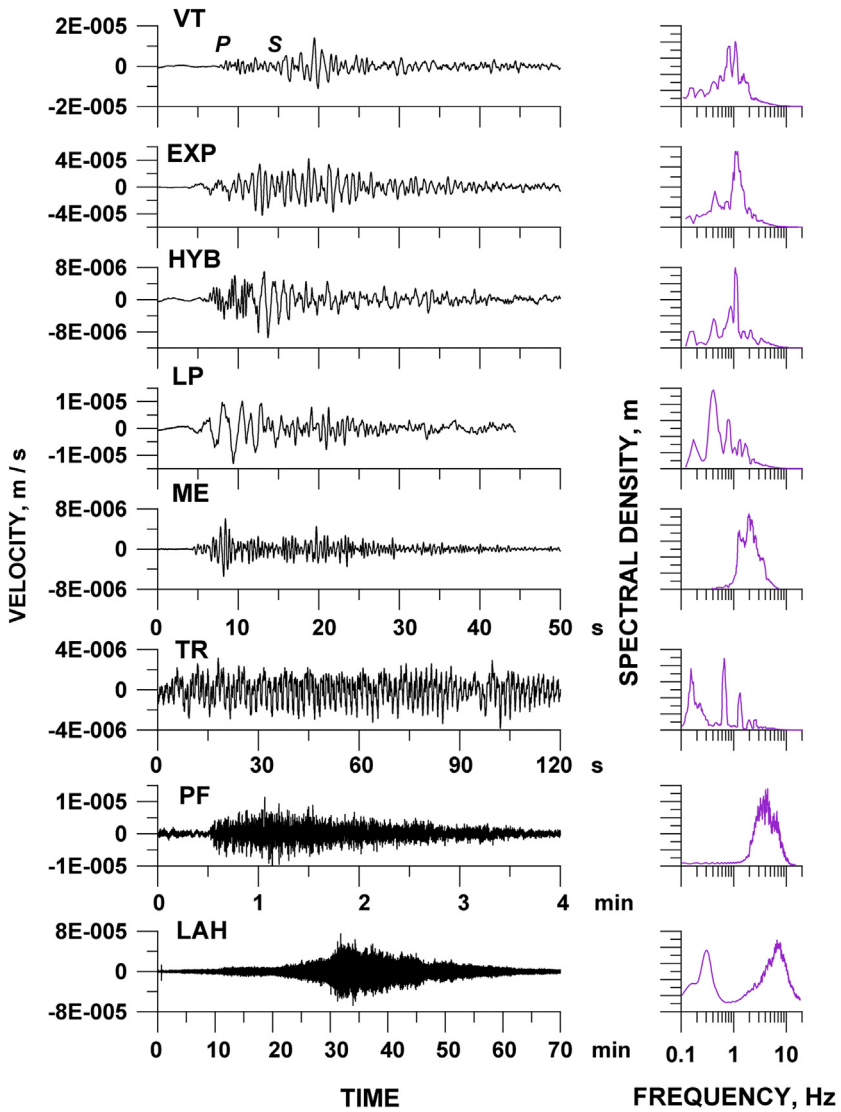


**FIGURE 2.6** Types of seismic signals (velocity, vertical component) recorded during the 1997–2015 unrest at andesitic Volcán de Colima (left side) and corresponding Fourier spectra (right side). VT, volcano-tectonic earthquakes; EXP, seismic signals associated with magmatic explosive events; HYB, hybrid signals associated mainly with white cloud degassing; LP, long-period signal recorded at different stages of unrest; ME, microearthquake signals, forming together with short hybrid signals the microearthquake swarms, associated with extrusive and explosive activity; TR, volcanic tremor recorded at different stages of unrest; PF, signals associated with pyroclastic flows or rockfalls of incandescent rocks; and LAH, signals associated with lahars. All seismic signals were recorded by the broadband seismic station at a distance of 4 km from the crater and corrected for instrument response. *Courtesy: Colima Volcano Observatory.*

microearthquake events were associated with the magma movement before lava extrusion and during different stages of strong explosive activity. The volcanic tremor signals were observed at different stages of eruptive activity without any regularity. The signals marked as PF and LAH were associated with the movement of pyroclastic flows and rockfalls of incandescent rocks and lahars, respectively.

Generally, these numerous types of volcanic earthquakes may be included within two groups. The first group includes earthquakes generated by stresses, which arise during magma migration in Earth's crust. They may be caused by shear or tensile failure and occur during magma movement from the depth to Earth's surface through conduits, dikes, and so on, producing rupture at the tip of the magma body and within adjacent fault systems. They are called volcano-tectonic earthquakes; they correspond to A- and B-types of [Fig. 2.4](#). Their study is a study of physics of magma movement to the surface. These earthquakes are very important for forecasting of volcanic eruptions.

The second group includes earthquakes originated due to magmatic activity within the volcanic conduit, during volcanic explosions, or associated with the lava dome growth, the movement of lava flows, pyroclastic flows, and lahars. These signals may form a composite waveform beginning with volcanic tremor and continuing with the seismic signal of explosion and pyroclastic flow or rockfall, and then again volcanic tremor, as was noted for volcanic earthquakes of Santiaguito volcano, Guatemala, by [Johnson et al. \(2000\)](#). They are called eruption earthquakes. Their study is a study of physics of volcanic eruptions.

## 2.3 SEQUENCES OF VOLCANIC EARTHQUAKES

A majority of volcanic earthquakes occur in sequences. [Mogi \(1963\)](#) described three main types of earthquake sequences ([Fig. 2.7](#)): mainshock–aftershocks sequence (Type 1), foreshocks–mainshock–aftershocks sequence (Type 2), and swarm (Type 3).

Earthquake swarms are generally defined as a sequence of events closely clustered in time and space without a single outstanding shock. In practice, the difference in magnitude between the largest and second largest events of a swarm is 0.5 magnitude units or less ([McNutt, 2000](#)). If this difference is greater, we have the mainshock–aftershock sequence (Type 1).

### 2.3.1 Sequences of Volcano-Tectonic Earthquakes

Volcano-tectonic earthquakes may occur in all the three types of sequences. The first two types of sequences are rare, and they occur prior to flank or fissure eruptions. The swarm sequences are the most common for volcanic earthquakes.

The Type 1 sequence was observed prior to the flank eruption of subglacial volcano Grímsvötn that occurred on September 30, 1996, in Iceland

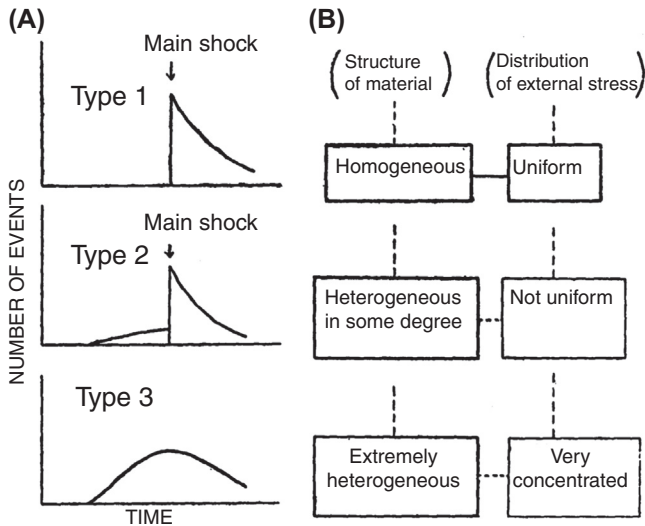


**FIGURE 2.7** Three main types of successive occurrences of elastic shocks sequences that were obtained from experimental destruction of material (A) and their relations to their structure of materials and the applied stresses (B). From Mogi, K., 1963. *Some discussions on aftershocks, foreshocks and earthquake swarms — the fracture of a semi-infinite body caused by an inner stress origin and its relation to the earthquake phenomena (third paper)*. *Bulletin of the Earthquake Research Institute, Tokyo University*, vol. 41, pp. 615–658.

(Zobin, 1999; Konstantinou et al., 2000). The mainshock with magnitude  $mb$  5.2 was the first in this earthquake sequence (Fig. 2.8A); the second largest earthquake occurred after about 3 hours with magnitude  $mb$  4.3 (ISC, 1996).

Seismic activity developed during three stages (Zobin, 1999). The first stage (first 19 hours) included the mainshock with magnitude  $mb$  5.2 ( $M_s$  5.4) and its aftershocks along the northern slope of the Bárðarbunga volcano, situated about 20 km to NW from Grimsvötn (Fig. 2.8B). Then, during the next 17 hours, the seismic foci marked a line connecting the two volcanoes, Bárðarbunga and Grimsvötn. This stage culminated in the opening of a fissure and the beginning of the eruption. The third stage was observed during the eruption (from midnight of September 30 till October 7) and included continuous seismic activity along the northern slope of the Bárðarbunga volcano as well as more western distributed epicenters, which together had clustered along the ring-fault epicentral zone.

The Type 2 sequence was observed before the January 1996 subaqueous phreatomagmatic eruption in the Akademia Nauk caldera (Karymsky Lake), Kamchatka. This eruption occurred after 4800 years of dormancy and resulted from the injection of juvenile basaltic magmas through a fissure (Fedotov, 1998). The preliminary seismic activity (Zobin and Levina, 1998)

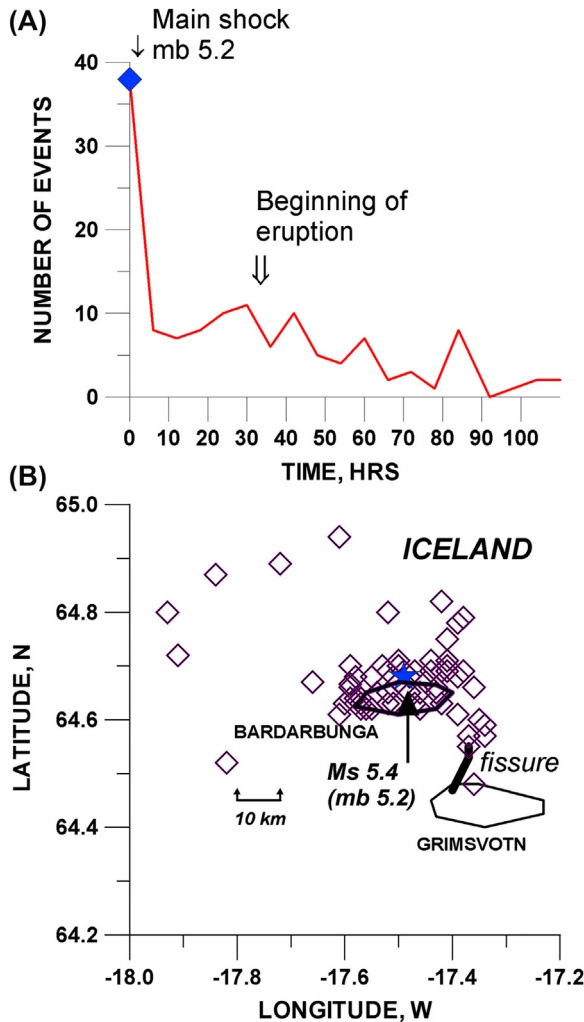


**FIGURE 2.8** Variations in 6-h numbers (A) and epicenter distribution (B) of the earthquakes associated with the September 30, 1996 subglacial flank eruption of Grimsvötn volcano, Iceland. The calderas of volcanoes Bárðarbunga and Grimsvötn are shown. The diamonds are epicenters; the star shows the epicenter of the mainshock. From Zobin, V.M., 1999. The fault nature of the  $M_s$  5.4 volcanic earthquake preceding the 1996 subglacial eruption of Grimsvötn volcano, Iceland. *Journal of Volcanology and Geothermal Research* 92, 349–358.

was represented by a foreshock–mainshock–aftershock sequence with a mainshock with magnitude  $M_w$  7.1 (Fig. 2.9).

The seismic foci of foreshocks migrated gradually along an NE–SW zone, including the forthcoming site of eruption and the mainshock  $M_w$  7.1 earthquake epicenter. The aftershocks spread to the southwest and northeast

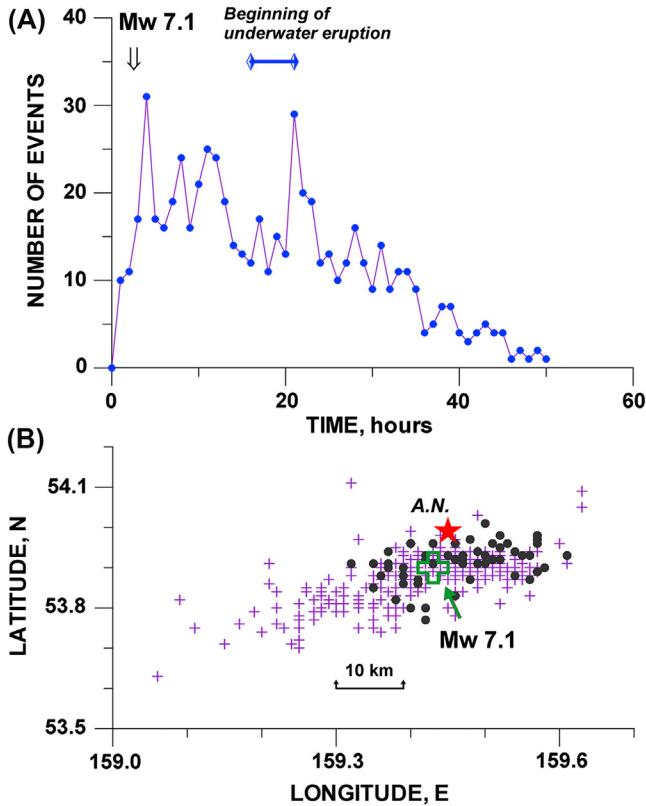


**FIGURE 2.9** Variation in hourly numbers (A) and epicenter distribution (B) of the earthquakes associated with the January 2, 1996 subaqueous phreatomagmatic eruption in Akademia Nauk caldera (Karymsky Lake), Kamchatka. The star shows the site of eruption in Akademia Nauk (AN) caldera; dots are the epicenters of foreshocks and crosses are the epicenters of aftershocks; large cross shows the epicenter of the mainshock Mw 7.1. The proposed time interval of the beginning of the eruption is shown. Data from *Kamchatka Regional Earthquake Catalog*.

from the mainshock, reaching the site of eruption; the total area occupied by aftershocks was of about  $50 \text{ km} \times 20 \text{ km}$ . The seismic foci were distributed within the depth range from 0 to 60 km. The eruption in Karymsky Lake began during a gradual decrease in the number of aftershocks.

The third type of earthquake sequence, and the most frequent type, is a swarm sequence. It occurs roughly with numerous earthquakes (Fig. 2.10). Fig. 2.11 shows the spatial–temporal characteristics of earthquake swarm associated with the September 1977 deflation of Krafla volcano, Iceland (Brandsdóttir and Einarsson, 1979). A small basaltic eruption broke out on a 0.9-km-long fissure the northern rim of caldera and continued for about 5 hours. The earthquake activity and volcanic tremor began about 2 hours before the eruption and accompanied the eruption.



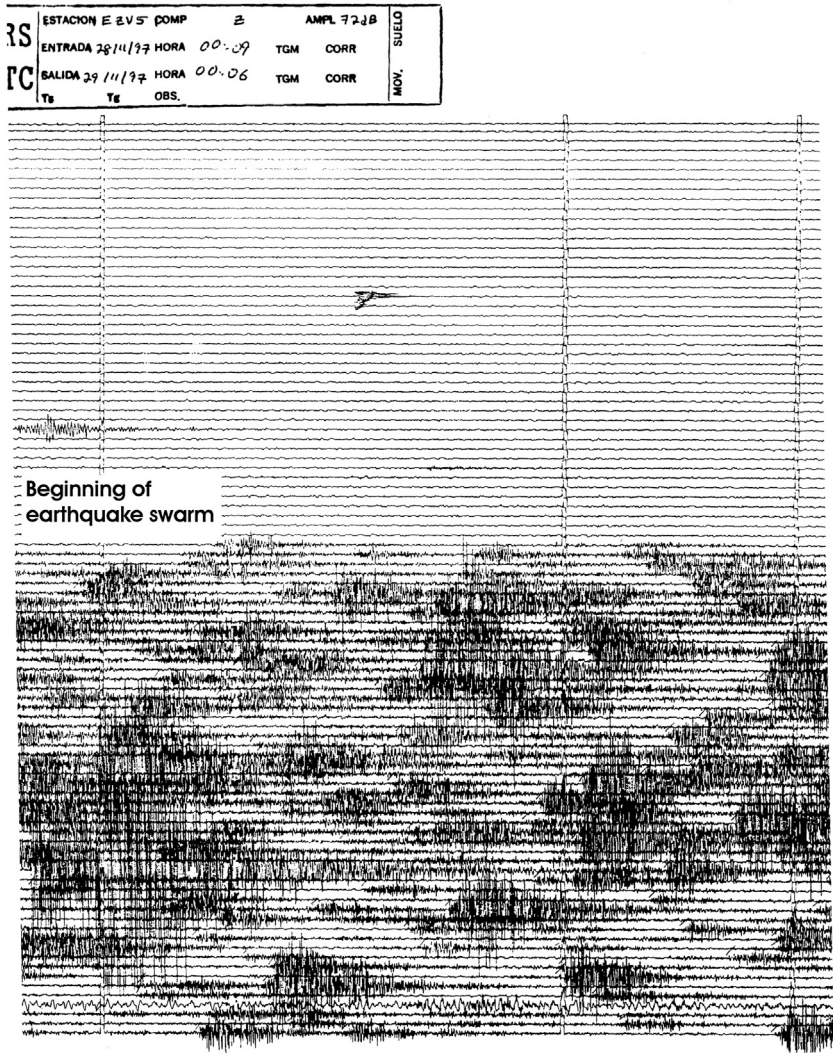


**FIGURE 2.10** The seismic signals representing a swarm of volcano-tectonic earthquakes at Volcán de Colima, Mexico, that marked the beginning of lava extrusion on November 28, 1997. The seismogram was recorded at a distance of 4 km from the crater. Minute marks are shown. *Courtesy: Colima Volcano Observatory.*

The first very small earthquakes were located in the central part of the caldera; later, the earthquake epicenters migrated southward with a rate of about 0.5 m/s. The seismic activity culminated just after the ceasing of lava eruption with eight earthquakes with magnitude 3 or more (Fig. 2.11). It is interesting to note that the 1138-m-deep well, situated along the way of earthquake migration, throws out the glowing pumice at the time of

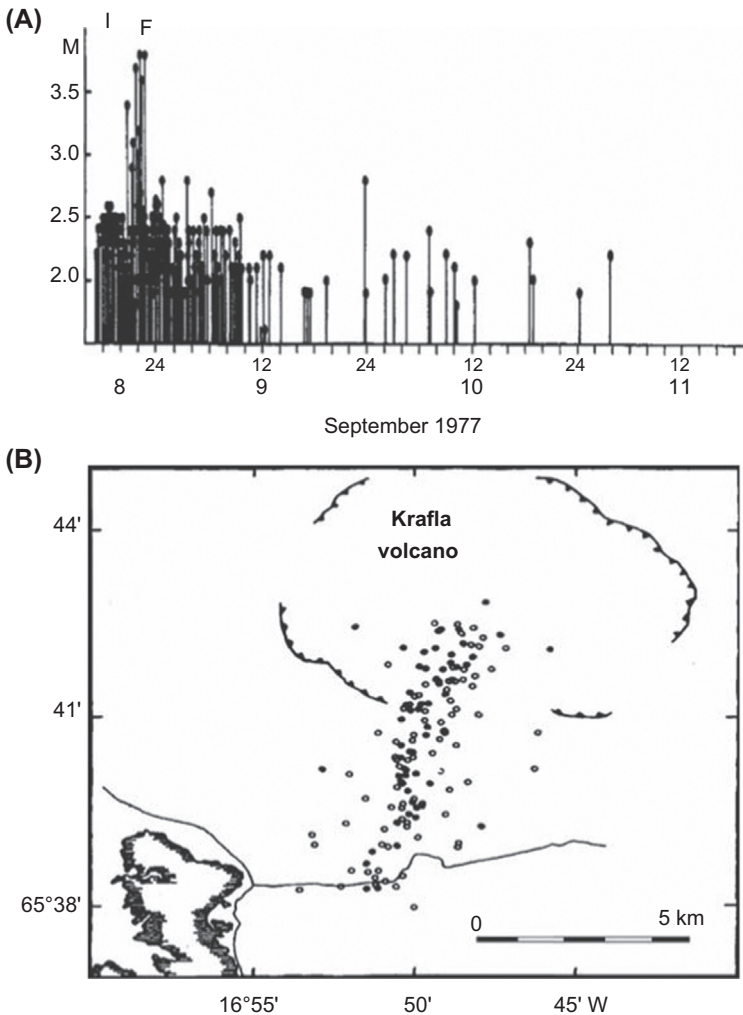


**FIGURE 2.11** The temporal variations of earthquake magnitudes in the swarm that occurred during the September 1977 deflation of Krafla volcano, Iceland (A) and the earthquake swarm epicenters (B). Symbol I marks the initial part of eruption; F marks the final part of eruption. A group of equal magnitude earthquakes in (A) is clearly seen. Dots are epicenters located with standard error less than 0.5 km; circles mark the epicenters of worse location. From Brandsdóttir, B., Einarsson, P., 1979. Seismic activity associated with the September 1977 deflation of the Krafla central volcano in northeastern Iceland. *Journal of Volcanology and Geothermal Research* 6, 197–212.

earthquake migration. It indicated that the earthquake foci migration was directly associated with the magma lateral migration.

Swarms may occur before the eruption, during the eruption, and after the eruption of volcano. Very frequently, the appearance of a swarm near a

volcano may indicate a forthcoming eruption of the volcano. Two main groups of volcano-tectonic earthquake swarms can be distinguished according to their characteristics of occurrence through time, their spatial position, and their relationship with volcanic activity (Gorelchik and Zobin, 1971).

*Group 1:* These swarms precede generally the summit eruptions of andesitic and dacitic volcanoes. The swarms consist of shallow volcanic earthquakes, the epicenters of which lie at a distance of 3–5 km from the crater, and the foci of which are distributed within the volcanic edifice at the depth range between 0 and 5 km above the sea level (a.s.l.). Depending on the deep structure of the feeding zones at andesitic volcanoes, earthquakes having focal depths of 10–15 km are feasible. The maximum magnitudes of earthquakes may reach 5–6. The swarm duration can be from 7–10 days to 3–5 weeks. During these swarms, the total energy, frequency, and maximum energy of earthquakes increase continuously and evenly, reaching their maximum values at the moment of eruption.

*Group 2:* These earthquake swarms precede generally flank eruptions at basaltic and andesitic volcanoes. Very shallow earthquakes predominate in the swarms. The epicenters of the earthquakes lie at a distance of 3–10 km from the place of formation of a new crater, and the foci are located at the depths of 0–5 km. The swarms preceding the flank eruptions at stratovolcanoes last for 1–10 days, and their maximum earthquake magnitudes reach 5–7. In these swarms, the number of events and their maximum energy sharply increase during the initial stage of the swarm, but, after reaching the peak frequency, begin to decrease gradually. The earthquakes cease a few hours before the flank fissure opening.

*Earthquake families:* The characteristic feature of volcano-tectonic earthquake swarms is the appearance of an earthquake family (a group of earthquakes having similar waveforms at a fixed seismic station). Kaufman and Burdick (1980) noted that during the swarm of volcano-tectonic earthquakes accompanied the collapse of Fernandina caldera, Galapagos Islands, in June 1968, the teleseismic records of large volcanic earthquake with magnitude  $M_s$  5.0–5.5 were spectacularly consistent from event to event (Fig. 2.12). The major seismic phases such as  $P$ -,  $PP$ -,  $SV$ -, and the Rayleigh-waves correlate closely. Even the fine details of the coda between the major phases match.

The families of small earthquakes, recorded at a distance of about 2 km, were founded during the dacitic dome growing at Usu volcano, Hokkaido (Okada et al., 1981; Okada, 1983). The waveforms in each family (Fig. 2.12) resemble each other in details from the onset to the coda, even in high frequencies ( $\geq 10$  Hz). Hypocenters of one family cluster in a very small region. Fig. 2.12 shows that the families OU1 and OU2 have completely different waveforms, although they are located less than 200 m apart.

Okada (1983) noted that about 80% of 2300 located volcanic earthquakes ( $M > 3.5$ ) occurred only in four families. Earthquakes of any particular family

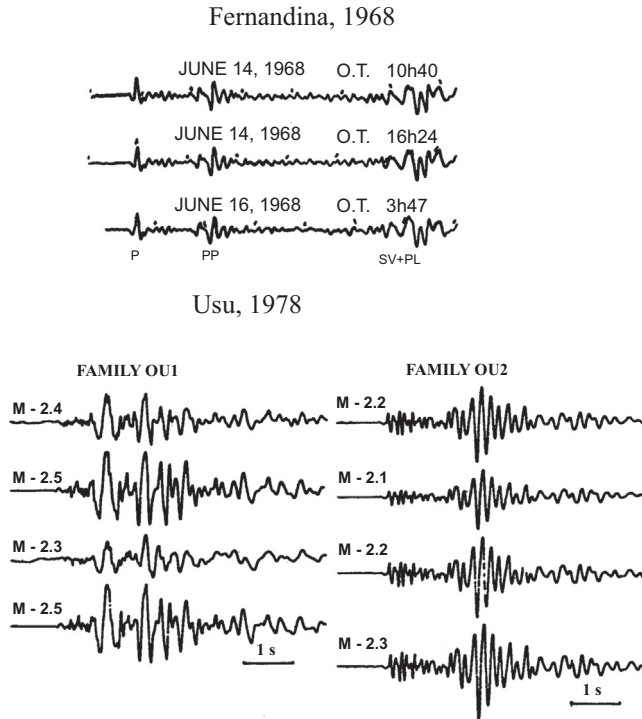


**FIGURE 2.12** Characteristic waveforms of volcano-tectonic earthquake families recorded at teleseismic distance during the caldera collapse of Fernandina volcano, Galapagos, in 1968 (long-period records) and at a near distance (about 2 km) during the dome growth at Usu volcano, Hokkaido, in 1978 (short-period records). For Fernandina records, the minute marks are shown. After Kaufman, K., Burdick, L.J., 1980. *The reproducing earthquakes of the Galapagos Islands. Bulletin of the Seismological Society of America* 70, 1759–1770 and Okada, H., Watanabe, H., Yamashita, H., Yokoyama, I., 1981. *Seismological significance of the 1977–1978 eruptions and the magma intrusion process of Usu volcano, Hokkaido. Journal of Volcanology and Geothermal Research* 9, 311–334.

tend to exist over a narrow range of their size distribution. The waveforms for family earthquakes vary only in the amplitude and not in the spectral shape.

### 2.3.2 Sequences of Eruption Earthquakes

Eruption earthquakes may also occur in sequences. Their clustering depends on the style of volcanic activity. The destruction of lava dome may cause a long-term sequence of seismic signals produced by the collapse of the lava body pieces and their movement along the volcanic flanks (Fig. 2.13). The permanent process of degassing can produce a sequence of explosion earthquakes (Fig. 2.14). The magmatic processes beneath the crater floor generate continuous volcanic tremor (Fig. 2.15). The nature of these sequences of



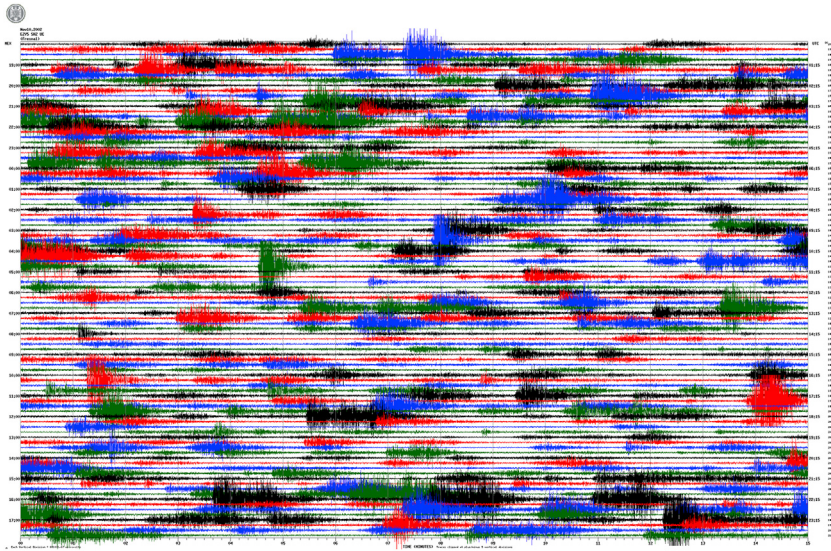


**FIGURE 2.13** Sequence of the seismic signals produced by rockfalls during the 2002 block-lava eruption at Volcán de Colima, Mexico. One-day short-period records at a distance of 1.7 km from the crater are shown. *Courtesy: Colima Volcano Observatory.*

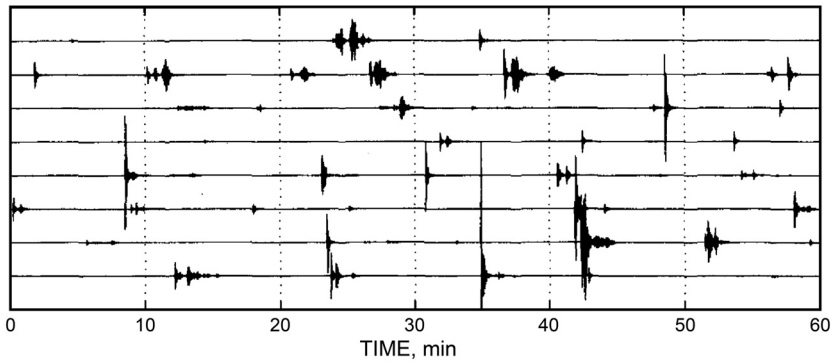


**FIGURE 2.14** Sequence of explosion earthquakes recorded at Karymsky volcano, Kamchatka, during the 1997 eruption. The 8-h broadband records at a distance of 1.5 km from the crater are shown. *From Johnson, J.B., Lees, J.M., 2000. Plugs and chugs – seismic and acoustic observations of degassing explosions at Karymsky, Russia and Sangay, Ecuador. Journal of Volcanology and Geothermal Research 101, 67–82.*

volcanic earthquakes differs from the nature of volcano-tectonic earthquake swarms and is related to the dynamics of deep and surface manifestations of magmatic activity.

These basic investigations had become the foundation for the construction of volcanic seismology as a science. During the last 30 years of the 20th

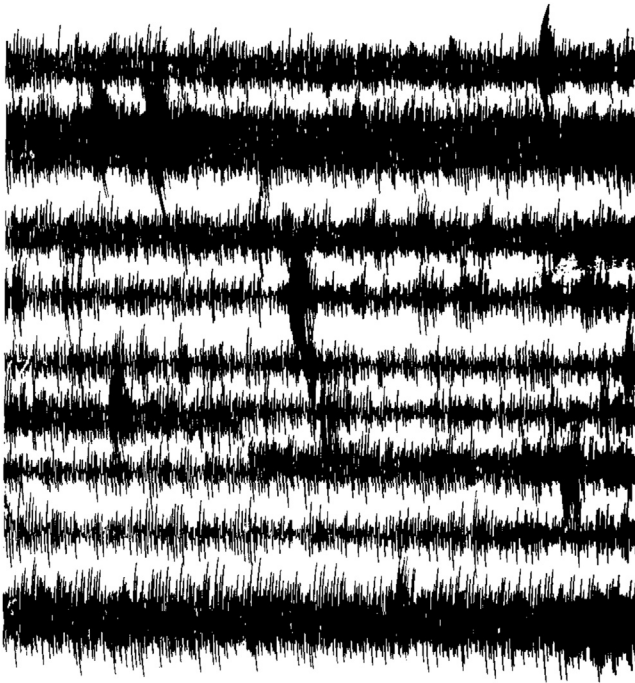


**FIGURE 2.15** Sequence of continuous volcanic tremors recorded by a short-period instrument during sustained lava eruption of Kilauea volcano, Hawaii. From Koyanagi, R., Chouet, B., Aki, K. (1987). *Origin of volcanic tremor in Hawaii. Part I.* In: Decker, R.W., Wright, T.L., Stauffer, P.H. (Eds.), *Volcanism in Hawaii*. U.S. Geological Survey Professional Paper, vol. 1350, pp. 1221–1257.

century and the first two decades of the current century, the detailed and complex studies of some eruptions (among them, New Tolbachik volcanoes, Kamchatka, 1975–1976; Mt. St. Helens, Cascades, 1980; 2004–2005; Redoubt, Alaska, 1989–1990; Pinatubo, Philippines, 1991; Soufrière Hills, Montserrat, 1995–2013; Volcán de Colima, México, 1998–2017; El Hierro, Canary Islands, 2011–2012) strongly increased our understanding of the seismo-volcanic process. It was shown that the source of the majority of A-type volcano-tectonic earthquakes is similar to the source of tectonic earthquakes (Zobin, 1970; 1972); it was shown that the records of explosive volcanic earthquakes can be interpreted as Lamb pulses excited by a nearly vertical single force that represents the counter force of the eruption (Kanamori and Given, 1983; Nishimura, 1998); it was shown that shallow earthquakes in volcanic areas often have mechanisms with isotropic components, indicating volume changes of either explosive or implosive polarity (Miller et al., 1998).

Many models have been proposed for explaining the nature of volcanic tremor (Steinberg and Steinberg, 1975; Aki et al., 1977; Schick, 1981;

Ripepe and Gordeev, 1999; etc.) and long-period volcanic earthquakes (Chouet et al., 1994; etc.). Of course, the methodology for volcanic eruption forecasting was developed (Tokarev, 1978; Malone et al., 1981; Harlow et al., 1996; Reyes-Dávila and De la Cruz-Reyna, 2002; etc.). Installing broadband digital equipment at volcanoes gives a good perspective for better understanding the nature of seismic signals originating with volcanic activity (Chouet, 1996).

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## Chapter 3

# Fundamentals of Volcanic Seismology

The Earth's volcanism is the manifestation at the surface of the planet of internal thermal processes through the emission at the surface of solid, liquid, or gaseous products (Francis, 1998). The variety of eruptive processes, related to magma movement at depth, and occurring during the emission at the surface of the solid, liquid, or gaseous products, may generate seismic signals. Therefore the fundamentals of volcanic seismology represent not only the mathematical approximation of the sources generating the seismic waves but also the description of the magmatic processes that can produce volcanic earthquakes and of the physical properties of the medium where the earthquakes occur. In this chapter, the author discusses the physical aspects of magma flow at depth, the experimental evidences of the generation of seismic signals at the different stages of magma flow and the tensor representation of seismovolcanic sources.

### 3.1 MAGMA FLOW WITHIN THE VOLCANIC CONDUIT

Volcanic eruptions represent the episodic or continuous surface discharge of magma from a storage region or a magma reservoir. The ascent of magma and its movement to the surface from the magma reservoir may be considered as steady or unsteady multiphase flow. For magma to erupt, the magma reservoir pressure has to exceed the sum of frictional and magmastatic pressure losses, as well as losses associated with the opening of the conduit. Fig. 3.1 shows a general scheme of magma ascent from the deep magma reservoir through the volcanic conduit. The magma reservoir contains melt, crystals, and dissolved gas. Magma is saturated in volatiles at the top of the reservoir. Bubbles nucleate at a slightly higher level (the exsolution surface), as magma rises and experiences a decrease in pressure. These bubbles are critical because they provide the driving force for eruptions by increasing either buoyancy or magma reservoir overpressure. The magma is a mixture of silicate melt and gas bubbles formed between the exsolution and fragmentation surfaces. Fragmentation occurs when gas occupies 70%–80% of the available volume. At this point, the magma transforms from a liquid with suspended bubbles to a gas with suspended clots of liquid. The abrupt density decrease (expansion)

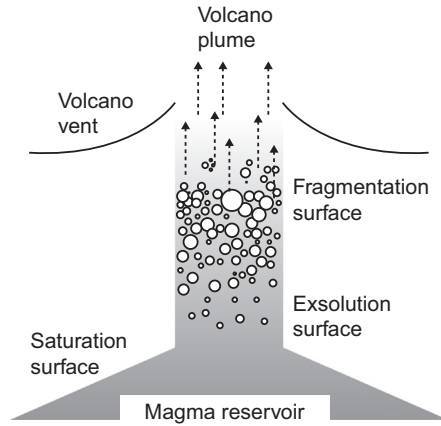


**FIGURE 3.1** Schematic illustration of processes that occur in volcanic conduits. From Cashman, K.V., Sturtevant, B., Papale, P., Navon, O., 2000. *Magmatic fragmentation*. In: Sigurdsson, H. (Ed.), *Encyclopedia of Volcanoes*. Academic Press, San Diego, pp. 421–430.

that accompanies this transformation results in the acceleration of the gas mixture out of the vent, in the form of a volcanic plume (Cashman et al., 2000; Gonnermann and Manga, 2012).

Seismic signals may be generated at any point in this system, but the type of seismicity depends on the physical processes and the state of the magmatic fluids at each stage of magma flow. The fluid-filled cavities and cracks within the volcanic conduit may be considered as two-phase solid–liquid systems and serve as the oscillators generating seismic waves. The magma ascent within the conduit before an explosion also generates seismic signals.

### 3.1.1 Magma Flow Regimes

Magma flow in the conduit is characterized by three regimes (Melnik et al., 2005; Gonnermann and Manga, 2007).

#### 3.1.1.1 Homogeneous Magma Flow

Homogeneous magma flow occurs in the lowest zone, between the magma reservoir and the exsolution (nucleation) surface of the conduit. The ambient pressure,  $p$ , within this zone is higher than the nucleation pressure,  $p_{\text{nuc}}$ , and the magma enters the conduit. For the given initial concentration of dissolved gas  $c_0$  and solubility coefficient  $k_c$  with  $(p > p_{\text{nuc}} = c_0^2/k_c^2 - \Delta p_{\text{nuc}})$ , a standard model of a viscous liquid can be applied.

#### 3.1.1.2 Bubbly Liquid Magma Flow

Bubbly liquid magma flow takes place in the intermediate zone, between the exsolution (nucleation) and fragmentation surfaces where  $p < p_{\text{nuc}}$ . Within

the nucleation region, the bubbles are formed with a number density  $N_d$ . Bubble nucleation requires a supersaturation pressure to overcome the energy barrier provided by surface tension. Supersaturation may be achieved if diffusion of volatiles from the melt into the bubbles cannot keep pace with the decrease in volatile solubility caused by ascent-driven decompression. Once bubbles nucleate, ascent-driven decompression continues to decrease volatile solubility, causing the volatile diffusion from the melt to the melt–vapor interface. Due to viscous resistance, the pressure in the growing bubble,  $p_g$ , decreases more slowly than the pressure in the surrounding melt,  $p_m$ . This can result in a large overpressure in the growing bubble  $\Delta p = p_g - p_m$ , providing the magma ascent rate and magma viscosity suitably high.

The bubble growth rate depends on the rate-limiting processes such as the diffusion of volatiles to the melt–vapor interface, the viscous flow of the melt, and the rate of change in solubility caused by decompression. The merging of smaller bubbles to form larger ones promotes the separation of gas from the melt. The coalescence of the bubbles depends on melt viscosity and is probably only important in low-viscosity magmas. Volcanic rocks typically contain a power-law or exponential distribution of bubble sizes. The exsolution of volatiles from the melt phase during open-system degassing may occur as the gas bubble migrates from deeper to shallow parts of magma, gas may flow from bubble to bubble through small apertures in the surrounding melt shells, or bubble walls may be ruptured during magma fragmentation, allowing gas to escape.

### 3.1.1.3 A Gas-Particle Dispersion Magma Flow Regime

A gas-particle dispersion magma flow occurs above a narrow region of fragmentation that separates a zone of high-density, high-viscous magma from a zone of low-density gas-particle dispersion, the resistance of which is determined by the turbulent viscosity of the gas phase and is negligibly small. When  $\Delta p$  exceeds a critical value, fragmentation of the bubbly media occurs. A competing process is the coalescence of the bubbles with the development of a permeable porous structure and outflow of gas from the magma through a system of interconnected pores. This process reduces gas pressure and can also lead to a collapse of the porosity to form dense magma. The total resistance of the conduit and the average weight of the mixture are determined by the position of the fragmentation region.

The actual process of fragmentation is probably different in magmas of different rheologies. The likely fragmentation mechanism in low-viscosity magmas, such as Hawaiian fire fountains and Strombolian bubble bursts, is fluid instability. The fragmentation of an already vesiculated magma occurs by sudden decompression and bubble rupture. This mechanism may be applied to the failure of volcanic domes, and may also describe explosive disruption of viscous melts during Vulcanian eruptions. Relative rate of

bubble growth, magma transport, and gas loss will control the style of eruptive behavior (explosive vs. effusive; [Cashman et al., 2000](#)).

The interaction of magma (or lava) with groundwater or surface water can lead to phreatomagmatic fragmentation. The depth of fragmentation is likely to be determined by the depth of the aquifer. Phreatomagmatic fragmentation results in a spectrum of eruptive activity ranging from passive quenching of the magma/lava to explosive ejection of ash ([Morrissey et al., 2000](#)).

### 3.1.2 Modeling of Magma Flow Regimes

[Gonnermann and Manga \(2007\)](#) performed the model calculations for conduit flow regimes (conduit radius is 25 m) for two end-member cases of bubble number densities  $N_d = 10^9/\text{m}^3$  and  $N_d = 10^{15}/\text{m}^3$ . The eruptive style (explosive or effusive) is largely controlled by the magma flow rate  $Q_m$ , which is modulated by processes within the volcanic conduit below the fragmentation surface. Bubble number density,  $N_d$ , controls the timescale of bubble growth and the ability of volatiles to exsolve efficiently. For  $Q_m < 10^6$  kg/s degassing is solubility limited. The rate of permeable gas flow through the vesicular magma eventually exceeds the rate of volatile exsolution so that significant outgassing occurs. Eventually, magma viscosity becomes sufficiently large for brittle shear deformation near the conduit walls. At  $Q_m > 10^6$  kg/s supersaturation leads to secondary bubble nucleation (labeled  $N$ ) and shifts to larger  $N_d$ . Upon further ascent, viscous heating near the conduit walls results in shear localization and prevents the occurrence of shear brecciation. Finally, overpressure results in fragmentation (labeled  $F$ ) at  $p_m \cong 10$  MPa, which is equivalent to several kilometers below the surface, because of low wall friction of the gaseous flow above the fragmentation depth. With an increase in magma flow rate  $Q_m$ , the style of eruption may change from effusive to explosive.

The characteristics of magma flow constrain the type of volcanic seismicity during an eruption. Modeling of different stages of the magma flow and the experimental study of rock rupturing at high pressure and high temperature allows the formulation of the conditions for seismic source generation.

## 3.2 EXPERIMENTAL STUDIES OF THE VOLCANIC PROCESSES AND THEIR APPLICATIONS FOR THE SEISMIC SOURCES

The experimental studies of the volcanic processes are divided into two groups: the brittle fracturing in flowing magma and the modeling of ascending magma within volcanic conduits. The first group of experiments explains the nature of volcano-tectonic, hybrid, and long-period earthquakes, while the second group of experiments explains the nature of volcanic tremors and the variety of seismic signals associated with explosive eruptions.

### 3.2.1 Experimental Grounds of the Brittle Fracturing in the Rocks at High Temperatures and High Pressure

#### 3.2.1.1 High-Temperature Fracturing of Volcanic Rocks

Seismogenic faulting is assumed to be confined to cool, brittle rocks, with an upper temperature limit of 600°C. Tuffen et al. (2003, 2008) deformed high-temperature silica-rich magmas under simulated volcanic conditions to test the hypothesis that high-temperature magma fracture is seismogenic. Uniaxial and triaxial deformation experiments are conducted on the samples of both glassy and crystalline lavas at temperatures up to 900°C. The acoustic emissions (AEs) recorded during experiments show that seismogenic rupture may occur in both crystal-rich and crystal-free silicic magmas at eruptive temperatures, extending the range of known conditions for seismogenic faulting. According to Tuffen et al. (2003), fracture occurs in magma if

$$\varepsilon' \eta > \tau^s \quad (3.1)$$

where  $\varepsilon'$ ,  $\eta$ , and  $\tau^s$  are the shear strain rate, magma viscosity, and shear strength of magma, respectively. Magma viscosity must be in the range  $10^9$ – $10^{14}$  Pa s for the fracture to occur, assuming typical magma shear strength of  $10^6$ – $10^7$  Pa and a plausible range of strain rates between  $10^{-2}$  and  $10^{-6}$  per second (Tuffen et al., 2003).

Fig. 3.2 demonstrates a representative AE waveform generated by high-temperature fracturing of rhyolitic obsidian and its power spectrum. Energy is predominantly in the range of 100–300 kHz. The onset is abrupt, and the waveform is similar to the waveforms that were recorded during the brittle failure of obsidian at room temperature. Postexperiment sample analysis showed that numerous predominantly axial cracks had formed, with curved, conchoidal surfaces, and local zones of cataclasis, where slip had occurred between major subparallel cracks (Fig. 3.2C, upper image). Scanning electron microscopy images of surfaces (Fig. 3.2C, lower image) show micrometer-scale hackle markings, typical of brittle glass fracture. This experiment shows the possibility of the generation of volcano-tectonic earthquakes within high-temperature rocks.

#### 3.2.1.2 High-Pressure Deformation of Dry and Saturated Water Basalt Samples

Benson et al. (2010) carried out deformation experiments in which dry and saturated water basalt from Mount Etna volcano was deformed and fractured at an effective confining pressure representative of conditions under a volcanic edifice. In the case of the saturated water experiment, a confining pressure of 60 MPa was used with a pore pressure of 20 MPa, whereas a confining pressure of 40 MPa was used in the dry experiment. An array of

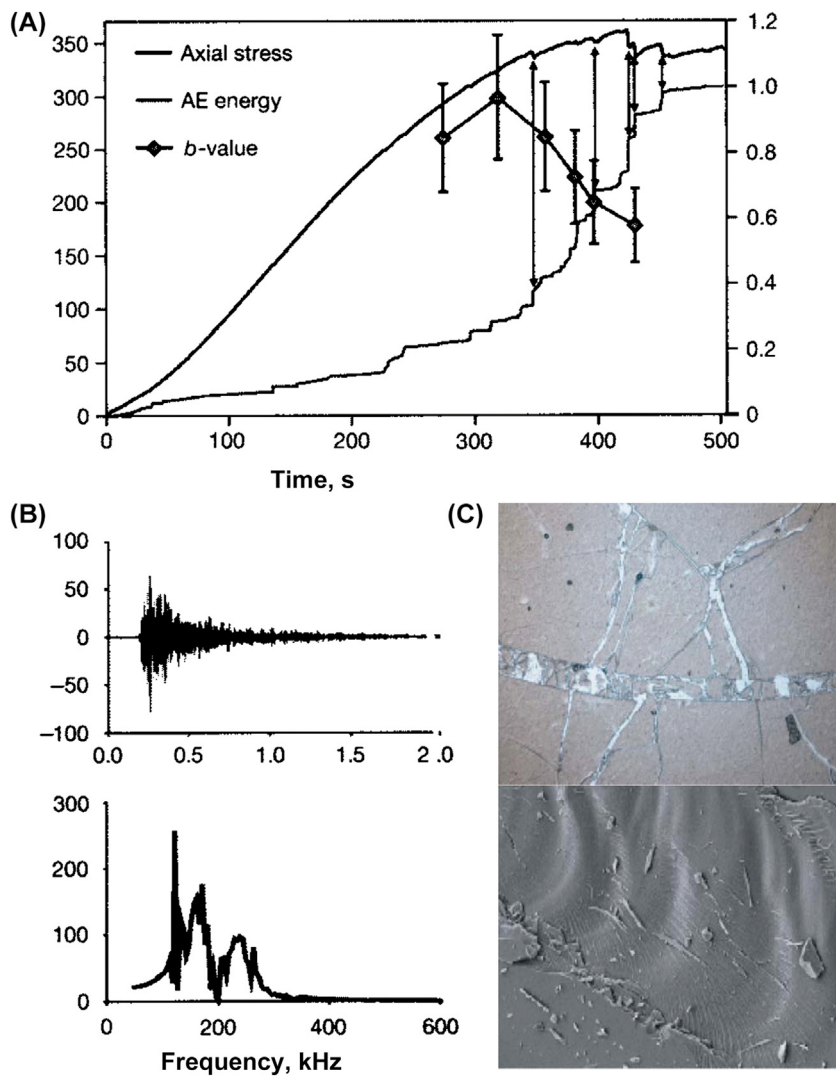


**FIGURE 3.2** Experimental results from high-temperature fracture of rhyolitic obsidian. (A) Axial stress, normalized cumulative acoustic emission (AE) energy, and AE  $b$  values against time for uniaxial deformation of rhyolitic obsidian at 645°C and 10–4.3 per second. Jumps in cumulative energy correspond to stress drops (arrows) and drops in compliance, indicating that cracking of the sample is associated with the release of acoustic energy. Error bars show 95% confidence limits. (B) Waveform (top) and power spectrum (bottom) of a typical AE event, showing the sharp onset and high-frequency content (predominantly 100–300 kHz) that are the characteristics of brittle failure. (C) Photomicrograph (top) of postexperiment obsidian sample, sectioned normal to the applied load, showing the formation of the gouge on curved brittle fracture surfaces; scanning electron microscopy image (bottom) showing detail of typical fracture surface. From Tuffen, H., Smith, R., Sammonds, P.R., 2008. Evidence for seismogenic fracture of silicic magma. *Nature* 453, 511–514.

18 piezoelectric sensors to detect AE was installed around the sample (12) and embedded in the upper and lower steel platens (3 per platen), of which 16 were used for AE recording. Waveforms recorded during the deformation of dry samples consisted of high-frequency events equivalent to volcano-tectonic earthquakes (Fig. 3.3A) while the waveforms received during deformation of the saturated water samples frequently exhibited hybrid-type waveforms, characterized by a high-frequency onset and a low-frequency component in the tail (Fig. 3.3B). Benson et al. (2010) suggest that hybrid seismic events are likely to be produced by the dual process of crack nucleation and deformation producing the high-frequency onset and, once these fluid pathways are created, fluid moving through the damage/crack network producing low-frequency resonance is seen in the coda of the waveforms.

A notable trend was observed in reducing instances of hybrid events leading up to the failure stage in the experiments, suggesting that the pore fluid present in the rock moves sufficiently quickly to provide a resonance, seen as a low-frequency coda. Fig. 3.3C shows a waveform of a purely low-frequency event from the same experiment with a saturated water sample. This was induced by venting the water pore fluid through the top part of the apparatus. It has the effect of isolating the high-frequency generation mechanism (microcracking) showing a clear low-frequency signal with significant power at low frequencies (50–100 kHz). Benson et al. (2010) concluded that hybrid seismic events are, in fact, a common type of volcanic seismic event with either high- or low-frequency events representing end members, and whose proportion depends on pore fluid being present in the rock type being deformed, as well as how close the rock is to failure.

### 3.2.2 Experimental Grounds of the Origin of Seismic Signals during the Magma Ascending within the Volcanic Conduit

Our knowledge about the seismic ability of magma ascending within volcanic conduits is rather approximate. Benson et al. (2014) presented the general view of a volcanic system illustrating the likely sources of the various signals commonly detected at volcanoes (Fig. 3.4). According to this scheme, a shallow zone (Zone A) where fracturing dominates is postulated to exist in the uppermost part of the edifice. The fracturing gives way to seismicity due to fluid resonances deeper in the edifice (Zone B). Deeper seismicity (Zone C) due to resonances within the conduit and fluid itself may be generated still deeper in the volcanic plumbing system.

Sahagian (2005) reviewed a set of models of conduit processes presented during the 2002 *Volcanic Conduit and Eruption Modeling* International workshop and summarized the parameters of ascending magma taken by



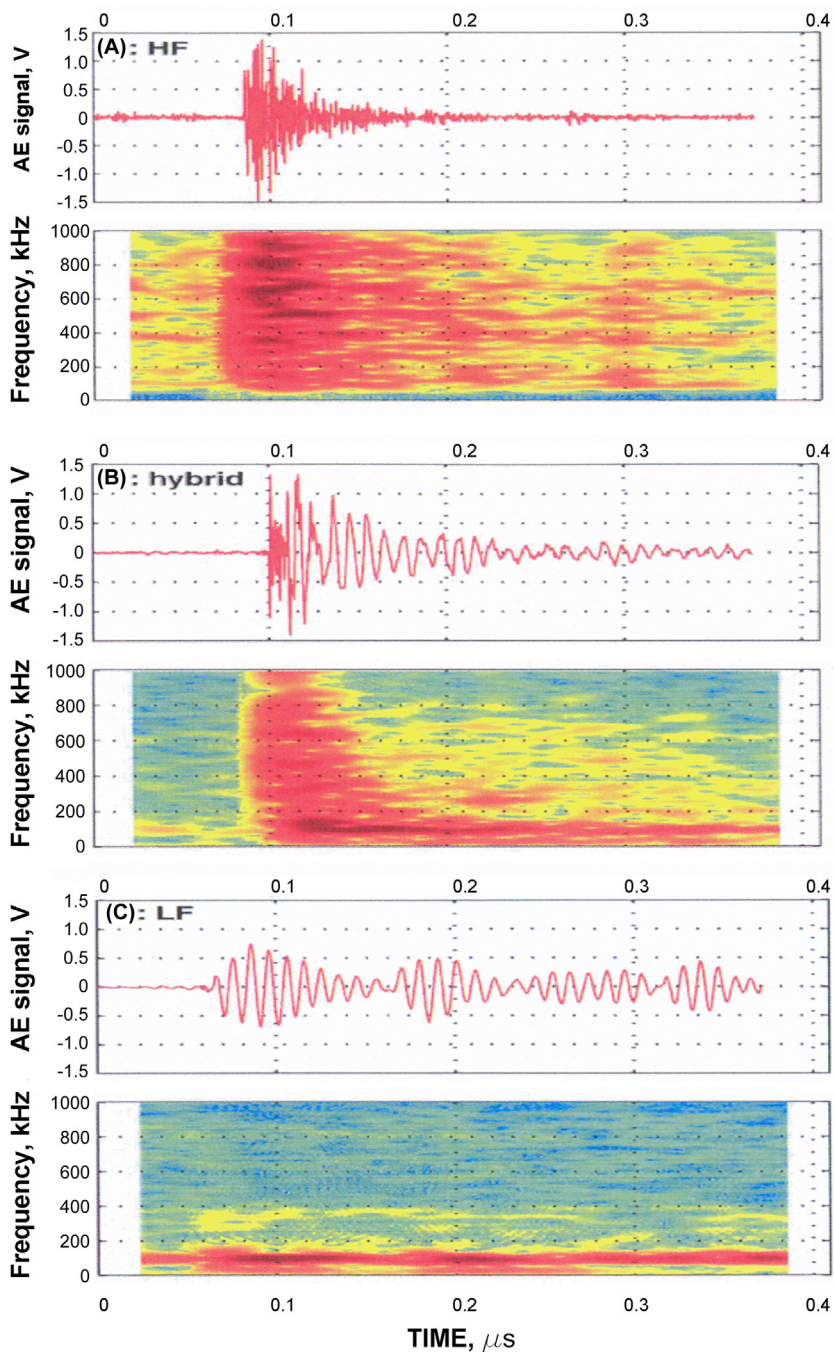


**FIGURE 3.3** Acoustic emission (AE) signals and their spectrograms obtained during deformation of a dry sample (high-frequency events, panel A), of a saturated water sample (hybrid events, panel B), and of a saturated water sample which was induced by venting the water pore fluid through the top part of the apparatus (low-frequency events, panel C). The top row in each panel illustrates representative waveforms, with power spectrograms plotted underneath (color denotes power). From [Benson et al. \(2010\)](#). *Spatio-temporal evolution of volcano seismicity: a laboratory study. Earth and Planetary Science Letters* 297, 315–323.

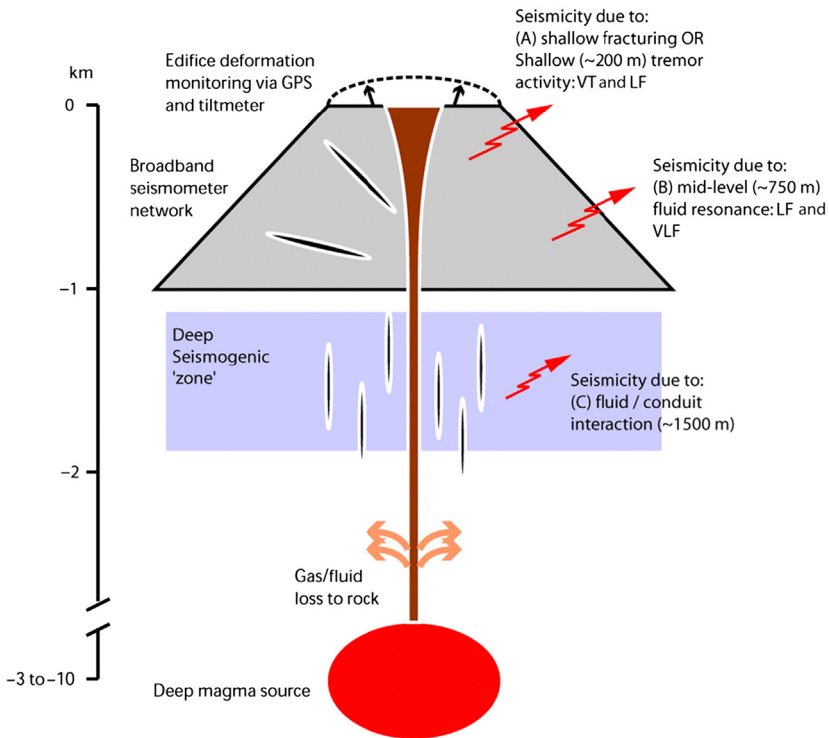


**FIGURE 3.4** A scheme illustrating the likely sources of the various signals commonly detected on the edifice. From [Cashman et al. \(2000\)](#). Magmatic fragmentation. In: “*Encyclopedia of Volcanoes*” (H. Sigurdsson, Ed.). Academic Press, San Diego, pp. 421–430.

researchers as the output parameters of their models. The summary of the quantities characterizing the processes of magma ascent is presented in [Table 3.1](#).

The author discusses two types of models: the models of explosive processes with rupturing of the surface of bubbly magma ([Alidibirov and Dingwell, 2000](#); [Lane et al., 2001](#)) and the models of explosive processes without rupturing of diaphragm to generate a gas-driven explosion ([Jaupart and Vergnolle, 1989](#); [Ripepe et al., 1996](#); [Chouet et al., 1997](#)).

### 3.2.2.1 Shock Tube Model

The shock tube model explains the explosive eruption by rapid decompression of magma in a shock tube ([Alidibirov and Dingwell, 2000](#); [Koyaguchi and Mitani, 2005](#)). According to this model, a bubbly magma at a high pressure is separated from air at the atmospheric pressure by a diaphragm. As the diaphragm is ruptured, a shock wave propagates into the air, and a

**TABLE 3.1** Generalized Model of Ascending Magma Parameters within Conduit.

| Parameter                                  | Quantity Range                             |
|--------------------------------------------|--------------------------------------------|
| Fragmentation Depth                        | 1080–5650 m                                |
| Fragmentation Pressure                     | 12–29 MPa                                  |
| Magma Gas Volume Fraction at Fragmentation | 0.60–0.83                                  |
| Exit Gas Volume Fraction at the Vent       | About 1                                    |
| Exit Dissolved Water Concentration         | 0.5–2.0 wt.% H <sub>2</sub> O              |
| Exit Pressure                              | 0.1–6.1 MPa                                |
| Discharge Rate                             | $0.7 \times 10^7$ – $7.7 \times 10^7$ kg/s |
| Exit Velocity                              | 118–182 m/s                                |

From [Sahagian \(2005\)](#). Volcanic eruption mechanisms: insights from intercomparison of models of conduit processes. *Journal of Volcanology and Geothermal Research* 143, 1–15.

rarefaction wave propagates into the bubbly magma. As a result, the bubbly magma is decompressed and expands. The magma fragments and the flow are assumed to be changed from bubbly flow to gas–pyroclast dispersion when the hoop stress or the gas volume fraction reaches a given threshold. Two types of fragmentation mechanisms are recognized: (1) high-viscosity magma fragments as the hoop stress reaches the tensile strength of the melt (stress fragmentation) and (2) the hoop stress does not grow in low-viscosity magma so that fragmentation occurs after bubble expansion when the gas volume fraction reaches a threshold (expansion fragmentation). This model places the explosion trigger at the magma surface.

[Arciniega-Ceballos et al. \(2014\)](#) modeled the microseismicity produced by volcanic explosions in a shock tube apparatus where the rapid depressurization of volcanic rocks was achieved. The opening of the diaphragms causes a pressure drop in the high-pressure autoclave, which yields a shock wave. Several experiments with samples with different porosity were performed under controlled pressure conditions (ranging from 4 to 20 MPa), at room temperature. The elastic energy related to the depressurization experiments was recorded using high-dynamic piezoelectric sensors. The experimental signals were associated with the explosive emission of gas or gas-particle mixtures caused by fragmentation of pumice rocks ([Fig. 3.5A](#)). The similarities between the experimental and volcanic signals produced by the degassing burst observed at Popocatepetl volcano, Mexico ([Fig. 3.5B](#)) suggest a similar physical process operating in both systems, experimental and volcanic.

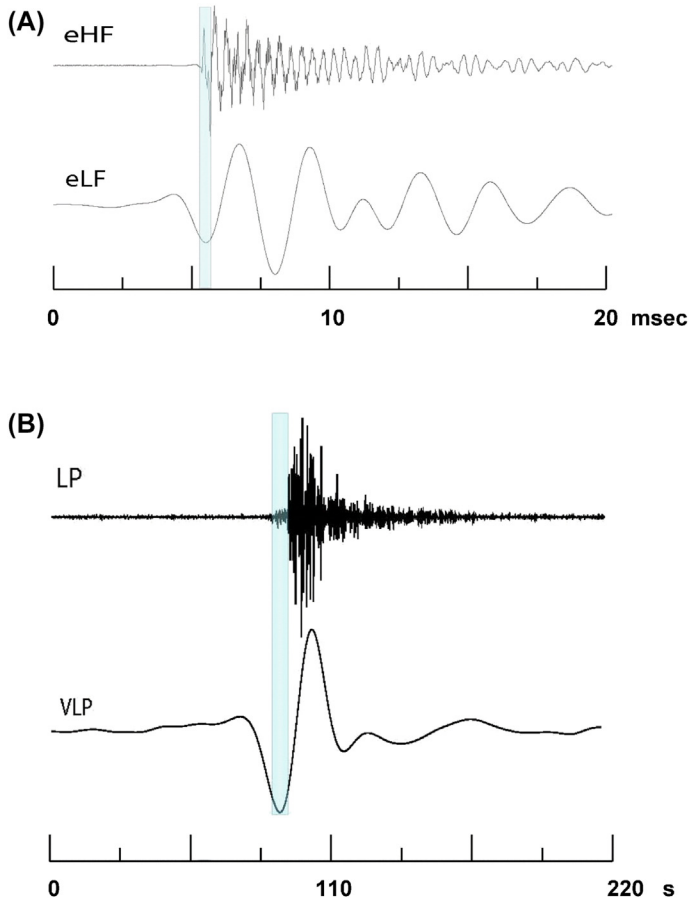


**FIGURE 3.5** Comparison of the experimental velocity records for a gas-burst experiment in a shock tube at 16.8 MPa using one diaphragm (A) and the seismic signals associated with degassing burst observed at Popocatepetl volcano, Mexico on January 19, 2000 at 01:14:31 GMT (B). In (A), the experimental high-frequency signal obtained after high-pass filtered experimental signals above 1 kHz, and its associated experimental low-frequency signal bandpass filtered in the band 20–800 Hz are shown. The experimental low-frequency signal starts milliseconds before the corresponding experimental high-frequency signal (marked with blue rectangle (light gray in print versions)). The vertical scale is relative to the maximum amplitude of the experimental high-frequency signal in arbitrary units. In (B), the original low-pressure signal and corresponding very low-pressure signal in the period band 15–100 seconds are shown. The high-frequency initial phase is marked with blue rectangle. The vertical scale is arbitrary. From Arciniega-Ceballos et al. (2014). *Seismological analysis of conduit dynamics in fragmentation experiments. Journal of Geophysical Research: Solid Earth* 119, doi:10.1002/2013JB010646.

### 3.2.2.2 *Modeling of Expanding Gas–Fluid Flows*

Expanding gas–liquid flows, designed to be analogous to those in volcanic conduits, were generated in a laboratory by Lane et al. (2001) using organic gas–gum rosin (natural pine resin) mixtures by expanding in a narrow vertical tube of borosilicate glass connected to a vacuum chamber. The tube was isolated from the vacuum chamber by a thin plastic diaphragm and an eruption was triggered by the rupture of the diaphragm. Depending on the level of volatile supersaturation, different flow regimes were observed, ranging from gentle unfragmented to violent fragmenting flows.

Fig. 3.6A shows the pressure traces obtained at various heights in the tube during a fragmenting diethyl ether-driven flow. Colors represent different flow regimes identified in video and still-camera images and in optical and pressure sensor data. Fig. 3.6B gives a schematic representation of the flow regimes and pressure oscillations in the fragmenting foam flow. Pressure oscillations are not detectable in region R4 (source liquid) despite oscillations being present in the flow regions above this region. The wave packets are dominantly present when the flow is fully detached from the tube wall but still interact with the wall on occasion. Some wave packets have a broadband signature and represent the impacts of gum rosin on the tube walls, while others display a series of narrowband peaks attributed to the resonant oscillation of the gas between foam slugs.

Pressure oscillations were detected over the whole range of flow conditions explored by Lane et al. (2001), and spectral analyses of these oscillations indicated that the dominant oscillation frequencies could be attributed to resonant oscillations in both foam and open gas spaces within the experimental tube. Flows, in which foam fragmentation occurred, displayed a broader, higher frequency, and more energetic spectrum than unfragmented flows.

Chouet (2003) noted that some of the pressure data obtained by Lane et al. (2001) show remarkable similarities to the seismic signals recorded for volcanoes. Fig. 3.7 compares the vertical ground displacement record of the very long-period event obtained at Kilauea volcano with pressure pulses recorded during the diethyl ether-driven flow near the optically detected fragmentation point. The Kilauea events occurred during a transition from slow deflation to rapid inflation of the volcano. Chouet (2003) noted that the reasons for these similarities are not yet clear, and suggested that similar fluid mechanical processes may be operated in both laboratory experiments and in the fluids moving beneath Kilauea.

### 3.2.2.3 *Modeling of the Effect of the Fluid Type on the Measured Signal*

Benson et al. (2014) designed an experiment generating a precisely controlled liquid-to-gas transition in a closed system, modeling the volcano conduit, by inducing rapid decompression of fluid-filled fault zones in a sample



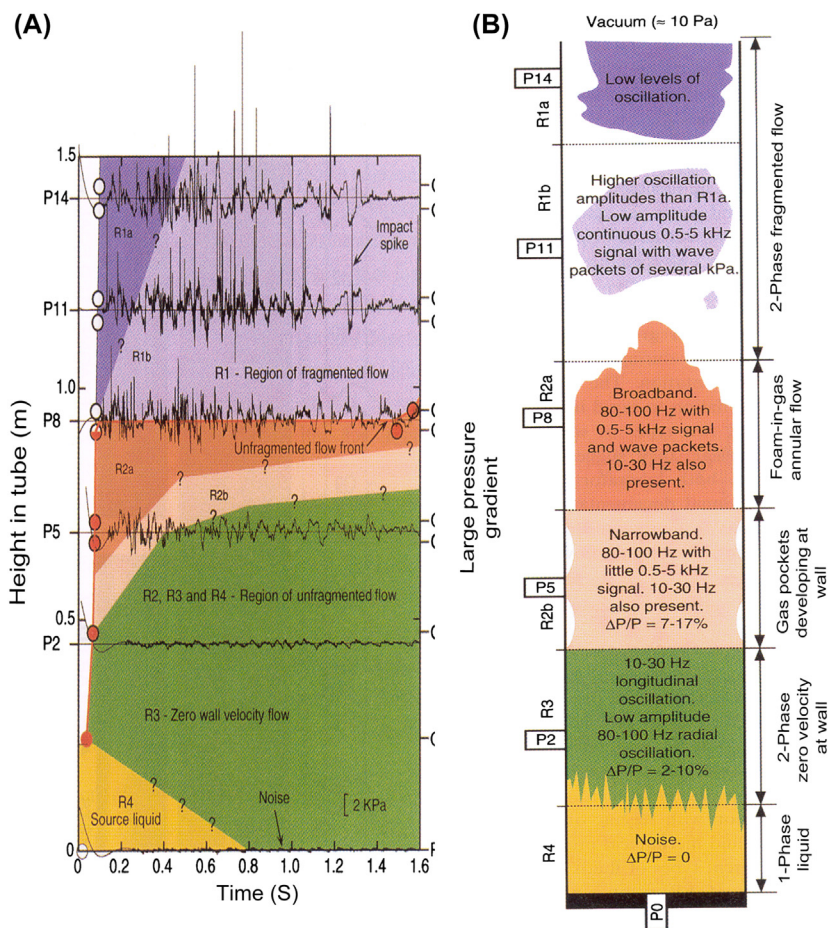


**FIGURE 3.6** Modeling of expanding gas–fluid flows and pressure oscillations in the fragmenting foam flow. (A) Representation of the fragmenting diethyl ether-driven flow in terms of height in the tube versus time plot of optical and pressure data. The positions of the pressure (P) and optical (O) transducers are indicated at the left and right, respectively. Open circles mark the times at which the first 270 foam fragments are detected, and filled circles indicate the onset of unfragmented flow. The circle at zero height represents the time at which P14 detected pressure falling below 99 kPa on decompression. The pressure data have been high-pass filtered at 5 Hz to remove transducer drift. The noise-free section of pressure data between 0.1 and 0.3 s is synthetic data included to reduce high-pass filter end effects. The pressure range (2 kPa) is indicated by the scale bar. Shading identifies regions in which different flow regimes are observed. (B) Schematic representation of flow regimes and pressure oscillations observed in a diethyl ether-driven fragmenting foam flow. The positions of the pressure transducers, indicated at the left and the flow regions, are representative of the conditions 0.3 s into the flow (A). From Lane et al. (2001). *Experimental observations of pressure oscillations and flow regimes in an analogue volcanic system. J. Geophys. Res.* 106, 6461–6476.

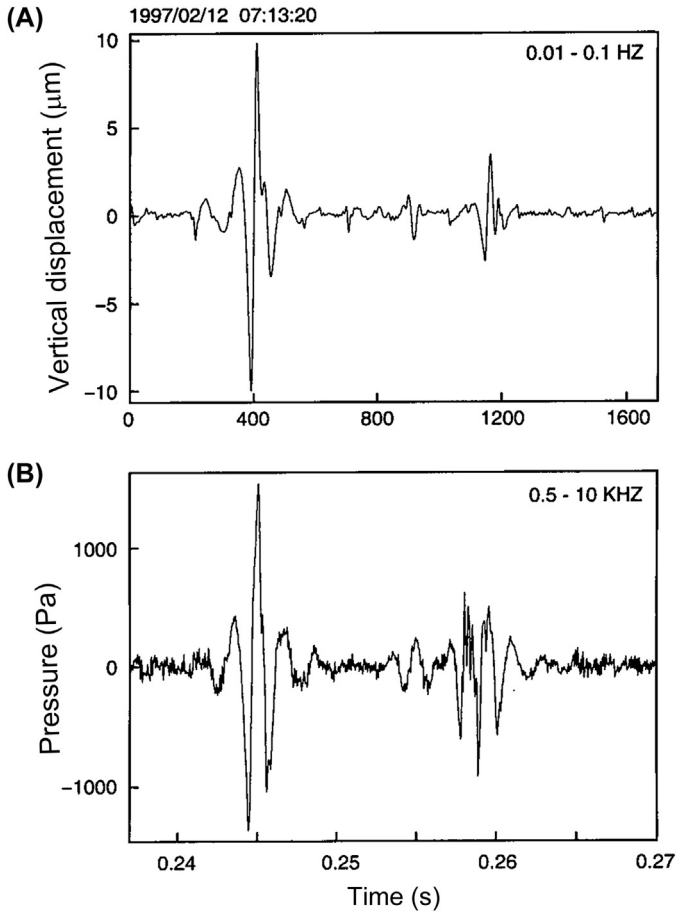


**FIGURE 3.7** Comparison between (A) very low-pressure displacement waveforms recorded at Kilauea volcano on February 12, 1997, during a transition from slow deflation to rapid inflation, and (B) pressure waveforms obtained at the boundary between regions R1 and R2 (see Fig. 3.4) in the diethyl ether-driven fragmenting foam flow. The record in (A) has been bandpass filtered in the 0.01–0.1 Hz band, and the record in (B) has been similarly filtered in the 0.5–10 kHz band. Date and time at the start of the Kilauea record are indicated at the upper left. Notice the different timescales. From Chouet (2003). *Volcano seismology. Pure and Applied Geophysics* 160, 739–788.

of basalt from Mount Etna volcano, Italy. To achieve the needed controlled pressure and temperature, the experiments used a triaxial testing apparatus fitted with an internal furnace capable of 200°C. The induced seismic activity was generated at pressure conditions equivalent to hydrostatic depths of 200–750 m. Benson et al. (2014) found that fluid phase transition is accompanied by a marked frequency shift in the accompanying microseismic data set that can be compared to volcano-seismic data.

Waveform data, obtained during the experiment with high-temperature (175°C) pore fluid (water), are shown in Fig. 3.8A, together with a time–frequency–power plot. As pressure decays over approximately 0.75 seconds decreasing through 2 MPa, a “swarm” of AE signals, is detected. During this sequence, two power peaks in the time–frequency spectra are seen, at 80–90 kHz and at approximately 20 kHz, corresponding to a transition of the low-frequency AE swarm to very low-frequency activity. Laboratory

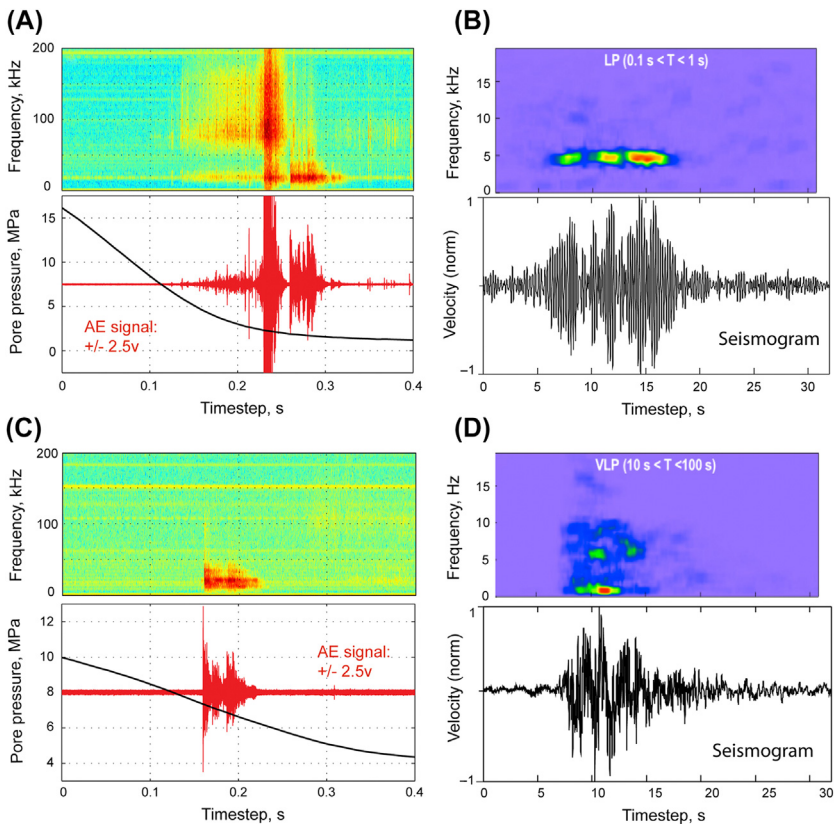


**FIGURE 3.8** (A) Results of an experiment with high-temperature (175°C) pore fluid (water). With pore pressure (black line) decrease through 2 MPa, the low-frequency acoustic emission (AE) swarm undergoes a transition to very low-frequency activity, seen as a switch from ~90 to ~20 kHz. (B) Low-frequency signal from Nisyros volcano. (C) Results of an experiment with dry nitrogen gas, also at 175°C. With pore pressure (black) decrease, only a single AE swarm is triggered at ~20 kHz and interpreted as a very low-frequency type signal. (D) Low-frequency signal from Nisyros volcano. The seismic signals were taken from Caliro et al. (2005). Note that the seismogram data have been normalized to maximum velocity in (B, D). From Benson et al. (2014). *Laboratory simulations of fluid/gas induced micro-earthquakes: application to volcano seismology. Frontiers of Earth Science* 2, 32.



data obtained may be qualitatively compared to low-frequency signals recorded at Nisyros volcano (Fig. 3.8B). The spectral peaks in Fig. 3.8A and B demonstrate a change in frequency from 90 kHz to 5 Hz or a factor of 18,000. It is comparable with a change in the scale of the fracture network from  $\sim 2$  cm scale to  $\sim 400$  m scale, a factor of approximately 20,000 which is in close agreement in terms of scaling.

Fig. 3.8C shows the results of the second experiment, conducted at similar conditions of confining pressure and temperature but with dry nitrogen gas as the pore fluid medium. In this case, as pore pressure decreases, only a single swarm of AE activity is seen at  $\sim 20$  kHz and interpreted as a very low-frequency type signal. Laboratory data obtained may be qualitatively compared to low-frequency signals recorded at Nisyros volcano (Fig. 3.8D). The observed peak frequencies of 10 kHz and 0.5 Hz are compared well to the change in length scale ( $\sim 2$  cm to  $\sim 400$  m).

### 3.2.2.4 Foam-Collapse Model

Foam is a mass of small bubbles formed in or on a liquid. Jaupart and Vergnolle (1989) proposed the process of the generation and collapse of a foam layer at the root of basaltic magma chamber to explain the different regimes of eruptions at basaltic volcanoes, such as fire fountaining at Kilauea or Strombolian explosions. To provide a unifying framework for these regimes, the phenomenon, induced by degassing in a reservoir that empties into a small conduit, was investigated. Laboratory experiments were conducted in a cylindrical tank topped by a thin vertical tube. Gas bubbles were generated at the tank bottom with the known bubble diameter and total gas flux. The bubbles rose through the tank and accumulated in a foam layer on the roof.

Depending on the behavior of the foam layer, three different regimes can be distinguished: (1) steady horizontal flow of the foam leading to bubbly flow in the conduit, (2) alternating regimes of foam build up and collapse leading to the eruption of a single, large gas pocket, and (3) flow of the foam partially coalesced into larger gas pockets leading to intermittent slug flow in the conduit. These regimes have natural counterparts in basaltic volcanoes.

Jaupart and Vergnolle (1989) showed that the necessary and sufficient condition for foam layer collapse is that the foam thickness  $h$  exceeds a critical value

$$h_c = \frac{2\sigma}{(\varepsilon\rho_1 g R)} \quad (3.2)$$

where  $\sigma$  is the coefficient of surface tension,  $\rho_1$  is the liquid density,  $g$  is the acceleration due to gravity,  $R$  is the bubble radius, and  $\varepsilon$  is the gas volume fraction in the foam. In a liquid of given physical properties, this occurs when the gas flux exceeds a critical value which depends on viscosity, surface tension, and bubble size.

The foam-collapse model places the explosion trigger at the initial deep source. The seismological application of the foam model was shown by [Ripepe et al. \(2001\)](#). They carried out a laboratory experiment, where bubbles in water were formed by air pumped at a constant flow rate from the bottom of a cylindrical tank of Plexiglas filled with water. They merged into a pipe imitating magma conduit. The gas foam accumulated at the roof of the tank and then collapsed into a large bubble inside the pipe. The bubble began to flow within the pipe and then finally broke at the liquid surface. During the movement of the bubble, an acoustic sensor in the tube outside the water recorded a low-frequency signal; when the bubble broke at the liquid surface, a high-frequency signal was recorded ([Fig. 3.9A](#)). The laboratory signals presented strong similarities to the seismic signal produced by a Stromboli explosion ([Fig. 3.9C](#)).

### 3.3 GENERAL DESCRIPTION OF THE SOURCE OF SEISMIC SIGNALS AT VOLCANOES

#### 3.3.1 Equivalent Force System Acting in the Earthquake Source

According to [Backus and Mulcahy \(1976\)](#), the seismic source of a tectonic earthquake is an indigenous source that originates within the Earth; its equivalent force system exerts neither total force nor total torque at any instance to the Earth and may be described completely by the seismic moment tensor. At the same time, the existence of seismic events at active volcanoes which can be described by a single-force equivalent system was demonstrated by [Kanamori and Given \(1982\)](#) and other seismologists. The displacement field generated by a seismic source at volcanoes is described by the representation theorem and may be written for a point source as follows ([Chouet, 1996](#)):

$$U_i(t) = F_j(t) * G_{ij}(t) + M_{jq}(t) * G_{ij,q}(t) \quad (3.3)$$

where  $U_i(t)$  is the  $i$ -th component of seismic displacement at the receiver at time  $t$ ,  $F_j(t)$  is the time history of the force applied in the  $j$ -th direction,  $M_{jq}(t)$  is the time history of the  $jq$ -th component ( $j, q = x, y, z$ ) of the moment tensor, and  $G_{ij}(t)$  is the Green's tensor function which relates the  $i$ -th component of displacement at the receiver position with the  $j$ -th component of impulsive force at the source position. The notation  $q$  indicates spatial differentiation with respect to the  $q$ -th coordinate and the symbol  $*$  denotes convolution. Summation over repeated indices is implied. Therefore the general source of seismic signals at volcanoes may be represented by a combination of its elements such as single force  $F$ , seismic moment  $M$ , and Green's delta function  $G_{ij}(t)$ . A short description of these three elements is explained in the following sections.

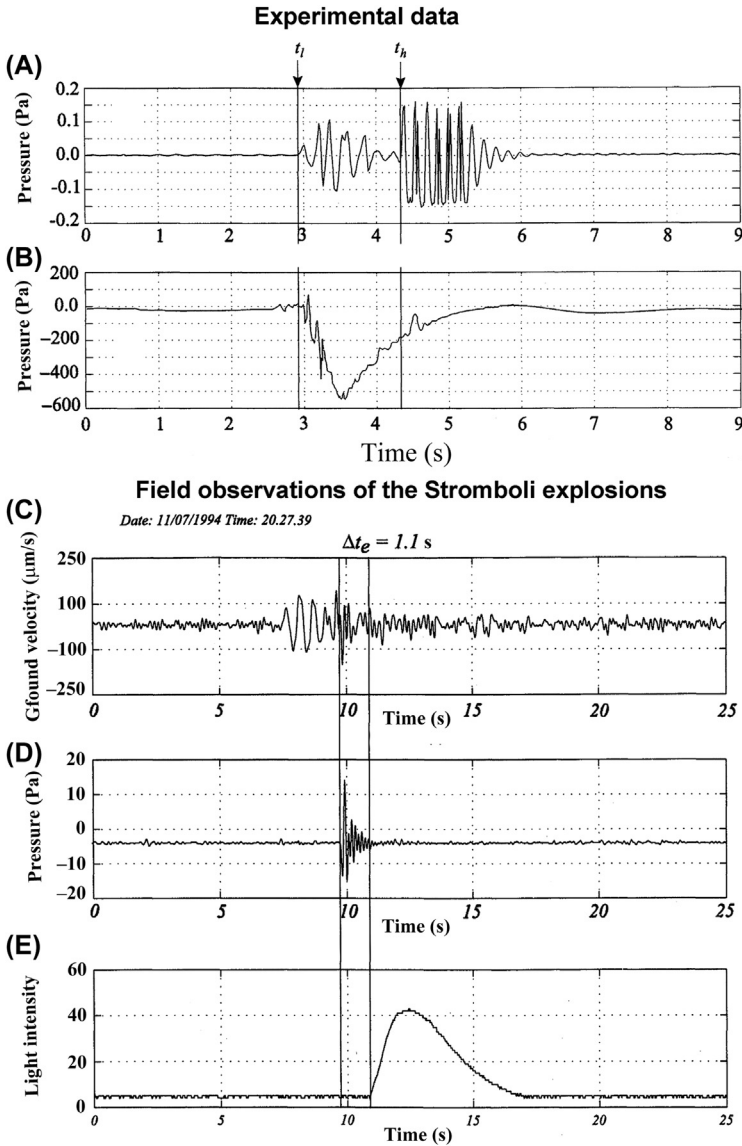


**FIGURE 3.9** Signals recorded during the laboratory experiments (A, B) and the seismic (C), infrasonic (D), and light-sensitive (E) recordings of the Stromboli explosion. (A) Low-frequency signal was recorded as soon as the gas slug started to rise ( $t_l$ ) followed by a high-frequency part ( $t_h$ ) as soon as the bubble film broke at the water surface. (B) Strong decompressive signal recorded by the sensor inside the water. Both signals are comparable to the signals recorded at Stromboli volcano (C, D). The time delay between seismic (C) and infrasonic (D) indicates that the source dynamics needs to trigger the explosion at the magma-free surface. The second time delay,  $\Delta t_e$ , between infrasonic (D) and light-sensitive (E) sensors depends on the gas jet velocity and the sound speed inside the conduit. From Ripepe et al. (2001). Time constraints for modeling source dynamics of volcanic explosions at Stromboli. *J. Geophys. Res.* 106, 8713–8727.

### 3.3.2 Green's Functions

Green's function  $G_{ij}(x, t; \eta, \tau)$  is the  $i$ -th component of displacement at  $(x, t)$  excited by the unit impulse applied at  $x = \eta$  and  $t = \tau$  in the  $j$ -th direction. Green's functions combine the elastic and inelastic effects of propagation from the source to the receiver and describe the signal that would arrive at the seismometer if the source time function was a delta function (Stein and Wysession, 2003).

### 3.3.3 Single Force

The force  $F_j$  applied at a particular point  $\xi_0$  is given as

$$F_j(\xi_0, \tau) = \iiint f_j(\eta, \tau) dV(\eta) \quad (3.4)$$

Takei and Kumazawa (1994) defined a single force  $F_j$  as a momentum exchange between the seismic source volume and the rest of the Earth. They stated that the single force originates from the difference between the density structure of the prescribed model and that of the actual value in the source region before the event and from the temporal change of the density structure in the source region caused by the finite displacement of mass during the event. The latter is a nonlinear effect caused by mass advection, which may occur by fluid flow, especially in volcanic regions. According to Chouet (2003), during the ascent of a slug of gas in a column of liquid toward the surface, the liquid moves downward to fill the void left behind by the ascending gas. The sinking of dense liquid changes the density structure of the fluid column and releases gravitational energy. Another example of a vertical single force is the recoil force generated by a volcanic jet during an eruption (Kanamori et al., 1984; Chouet et al., 1997). The single-force source may be used for the study of volcanic explosions and landslides (Kanamori and Given, 1982; Nishimura, 1995).

### 3.3.4 Seismic Moment Tensor

The symmetric second-order seismic moment tensor  $\mathbf{M}_{ij}$  describes the overall feature of the seismic source and consists of six independent components:

$$M_{ij} = \begin{pmatrix} M_{11} & M_{12} & M_{13} \\ M_{21} & M_{22} & M_{23} \\ M_{31} & M_{32} & M_{33} \end{pmatrix} \quad (3.5)$$

Each component of  $\mathbf{M}_{ij}$  corresponds to one set of opposing forces (Fig. 3.10). The seismic moment tensor, as a symmetric second-order tensor, has six independent components while the double-couple equivalent body force for a shear dislocation has only four degrees of freedom.

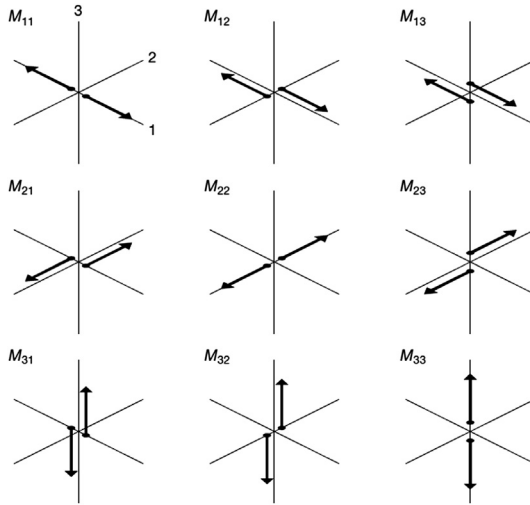


**FIGURE 3.10** Nine force couples corresponding to moment-tensor components. From [Kawakatsu and Yamamoto \(2015\)](#). *Volcano Seismology. Treatise on Geophysics 4, second ed.*, pp. 389–420.

These two extra degrees of freedom are the components of non-double-couple sources which may be considered as spherical, cylindrical, or tensile crack sources taken as its end members ([Julian et al., 1998](#); [Kawakatsu and Yamamoto, 2015](#)).

A double-couple source is applied for the description of the source process of volcano-tectonic earthquakes. Non-double-couple sources may be used for the study of volcanic explosions, volcanic tremors, and low-frequency earthquakes ([Aki et al., 1977](#); [Crosson and Bame, 1985](#); [Chouet, 1986](#); [Jousset et al., 2003](#)), as well as other unusual volcano-tectonic earthquakes ([Ekström, 1994](#); [Panza and Saraó, 2000](#)). For non-double-couple sources that represent the reservoirs of different forms filled by volcanic fluid, the amplitude ratios of three eigenvectors of the seismic moment tensor  $\mathbf{M}_{ij}$  can serve as criteria for discrimination between spherical, cylindrical, or tensile crack sources. [Table 3.2](#) presents the ratios of  $M_{11}:M_{22}:M_{33}$  calculated for different relationships between the elastic modulus (Lamè coefficients  $\lambda$  and  $\mu$ ). Normally, it is considered for rocks that  $\lambda = \mu$ ; however, for rocks at or near liquidus temperatures, the Lamè coefficients would be as  $\lambda = 2\mu$  ([Murase and McBirney, 1973](#)).

Therefore a general description of a volcanic seismic source is comprised of a combination of six seismic moment-tensor components and three single-force components for a point source ([Kumagai, 2009](#)). Mathematical aspects of seismic sources at volcanoes are given in the review papers by [Kawakatsu and Yamamoto \(2015\)](#) and [Kumagai \(2009\)](#).

**TABLE 3.2** Ratios of Three Diagonal Terms of the Seismic Moment Tensors  $M_{11}$ ,  $M_{22}$ , and  $M_{33}$  for Some Non-Double-Couple Sources of Volcanic Earthquakes (Kumagai, 2009; Kawakatsu and Yamamoto, 2015) for Different Relations between Elastic Modulus (Lamè Coefficients  $\lambda$  and  $\mu$ ).

| Type of Source           | $M_{11}:M_{22}:M_{33} (\lambda = \mu)$ | $M_{11}:M_{22}:M_{33} (\lambda = 2\mu)$ |
|--------------------------|----------------------------------------|-----------------------------------------|
| Spherical                | 1:1:1                                  | 1:1:1                                   |
| Vertical Cylinder        | 2:2:1                                  | 3:3:2                                   |
| Vertical Tensile Crack   | 3:1:1                                  | 4:1:1                                   |
| Horizontal Tensile Crack | 1:1:3                                  | 1:1:2                                   |

*Note.* If the crack is inclined, the ratios  $M_{11}:M_{22}:M_{33}$  are slightly different from the values proposed in the table.

### 3.3.5 Waveform Inversion

The seismic source parameters of volcanic earthquakes may be determined from waveform inversion by finding the best fit between observed and synthetic seismic waveforms. Waveform inversion of seismic signals of volcanic earthquakes is a more complicated task than the inversion of a tectonic earthquake. While the source of a tectonic earthquake is represented by the moment tensor corresponding to double-couple, seismic signals at volcanoes, as was noted earlier, may be generated by complex source processes, requiring a general source representation consisting of the single force and the moment tensor. Furthermore, instead of a step-like function for the source time function of a tectonic earthquake, the complex time history at the source of volcano-seismic signals, represented by Green’s functions, involves more unknowns.

The use of appropriate Green’s functions is critically important to obtain the correct solution in either the time or frequency domains. If the source is located in a medium that can be approximated by a homogeneous or layered structure with a flat surface, Green’s functions can be calculated by using propagator matrices and the discrete wavenumber method. However, this situation is not common in volcanic regions, and it is necessary to take into account steep topographies and the heterogeneous structure of volcanoes. The boundary integral equation method and boundary element method can be used to quantify the effect of three-dimensional topography in a homogeneous or layered structure. To treat both topography and structural heterogeneity simultaneously, the finite-difference method and the discrete lattice method are suitable (Kumagai, 2009).

The time domain waveform inversion of the seismic signals associated with the September 22, 1997, Strombolian explosion observed at Stromboli volcano, Italy, was performed by Chouet et al. (2003). Broadband velocity seismic records of 13 digital seismic stations were used in the inversion; the stations nearest to the crater are shown in Fig. 3.11. The station configuration gives a good coverage of the volcanic crater. The seismic record of the event under study, identified in the gray box, is shown in Fig. 3.12A. Before inversion, the seismic record was bandpass (2–30 seconds) filtered (Fig. 3.12B). According to the analysis of particle motions, the source was positioned within the small region northwest of the crater. In the inversion, six moment-tensor components and three single-force components for the point source were assumed. Green's functions were calculated by the finite-difference method using the topography of the volcano. Fig. 3.13 demonstrates a close match between the calculated and observed waveforms.

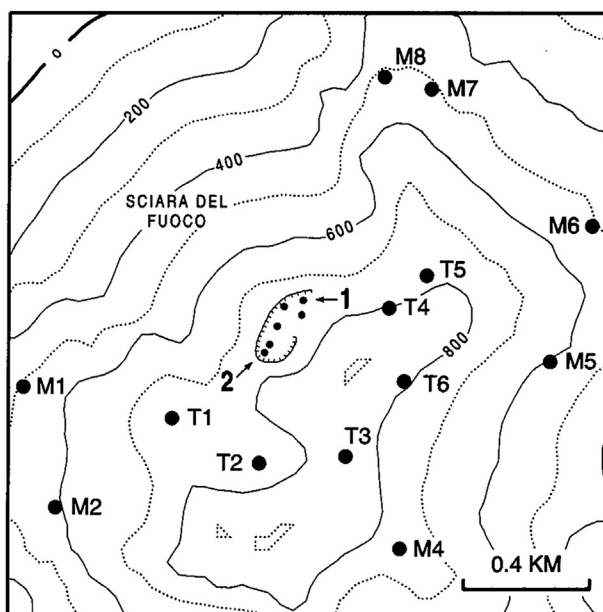


**FIGURE 3.11** Detailed map of summit area of Stromboli volcano, Italy, showing broadband seismic stations. (solid dots) in relation to crater (ticked line) and eruptive vents (small dots) Stations prefixed by “T” denote those of the “T ring” of sensors, and by “M” those of the “M ring,” located at the top and of the volcano, respectively. The arrows point to the locations of the two active vents at the time of the experiment in September 1997. Contour lines represent 100 m contour intervals. From Chouet et al. (2003). *Source mechanisms of explosions at Stromboli volcano, Italy, determined from moment-tensor inversions of very-long-period data. J. Geophys. Res.* 108, 2019.

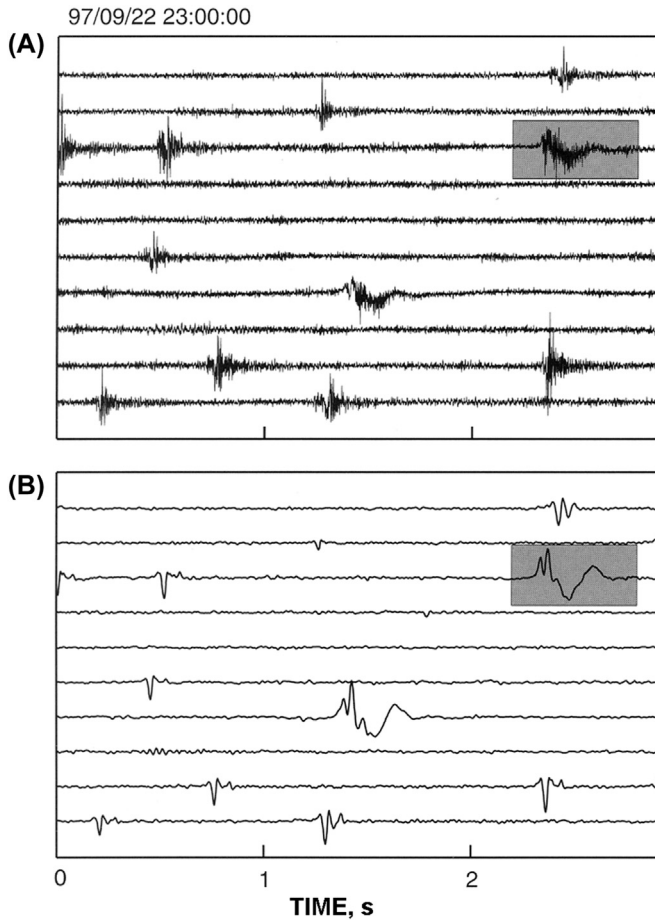


**FIGURE 3.12** Seismic record (A) of the east component of velocity at station T6 (for location, see Fig. 3.11) and the same record obtained by bandpass filtering the data in (A) in the 2–30 s band (B). The event under study is shown in gray box. From Chouet et al. (2003). *Source mechanisms of explosions at Stromboli volcano, Italy, determined from moment-tensor inversions of very-long-period data. J. Geophys. Res.* 108, 2019.

The result of this inversion is shown in Fig. 3.14. The volumetric components of the moment tensor clearly dominate. The amplitude ratio of the diagonal axes of the moment tensor is (1:1:2) which corresponds to a horizontal tensile crack source, considering that  $\lambda = 2\mu$  for rocks at or near liquidus temperatures (Table 3.2).

Volumetric source components are accompanied by a dominantly vertical force. The force amplitude, shown in Fig. 3.14, contributes roughly 13% of the signal amplitude relative to the contribution from the moment tensor.



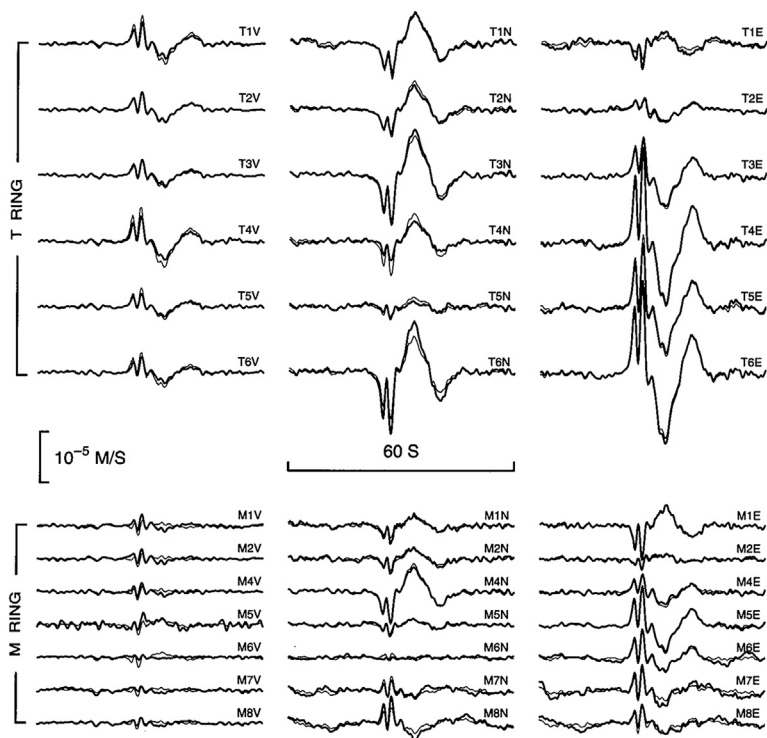


**FIGURE 3.13** Waveform match obtained for the September 22, 1997 explosive event, in which six moment-tensor components and three single-force components are assumed for the source mechanism. Thin lines indicate synthetics, and bold lines represent observed velocity waveforms. The station code and component of motion are indicated at the upper right of each seismogram. From Chouet et al. (2003). *Source mechanisms of explosions at Stromboli volcano, Italy, determined from moment-tensor inversions of very-long-period data. J. Geophys. Res.* 108, 2019.

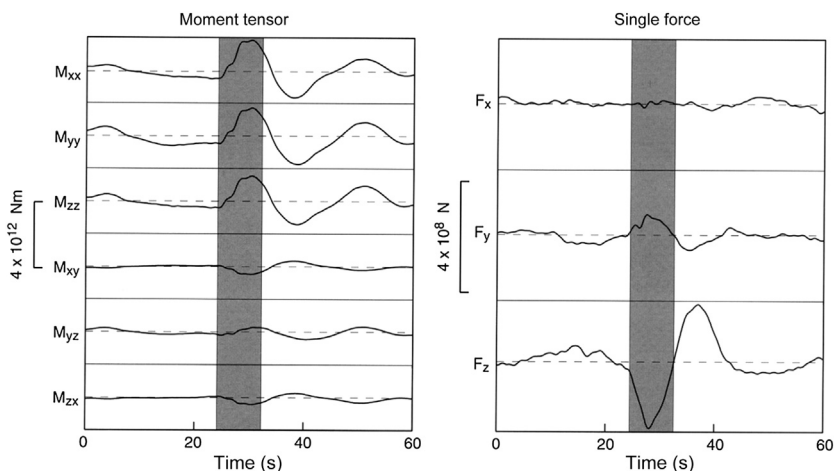


**FIGURE 3.14** Source time functions obtained for the September 22, 1997 event, in which six moment-tensor components and three single-force components are assumed for the source mechanism. Shading marks the interval during which the initial volumetric expansion of the source occurs. From Chouet et al. (2003). *Source mechanisms of explosions at Stromboli volcano, Italy, determined from moment-tensor inversions of very-long-period data. J. Geophys. Res.* 108, 2019.

The source centroid was located at an elevation of 480 m, 190 m north–northwest of Vent 2 (Fig. 3.15).

The seismological, volcanological, and experimental fundamentals of volcanic seismology will be used in the following chapters to develop models for the generation of seismic signals at different stages of volcanic activity.

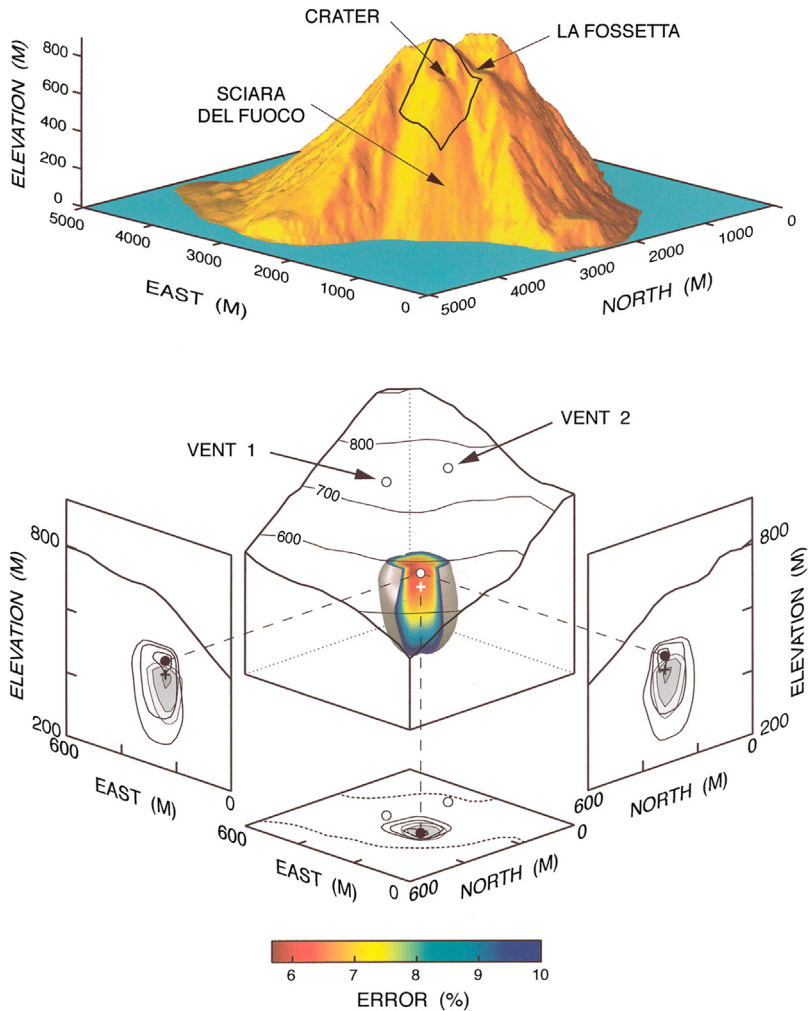


**FIGURE 3.15** Source location of the September 22, 1997 explosive event, occurring at Stromboli volcano. A southeast view of the Stromboli is shown at the top to help identify the location of the sources within the volcanic edifice. The bold black line on the northwest flank of the edifice bounds the surficial extent of the domain considered in the bottom view. Contours represent 6%, 8%, and 10% errors. From Chouet et al. (2003). *Source mechanisms of explosions at Stromboli volcano, Italy, determined from moment-tensor inversions of very-long-period data. J. Geophys. Res.* 108, 2019.

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## Chapter 4

# Origin of Volcano-Tectonic Earthquakes

In this chapter, the author considers a group of volcano-tectonic earthquakes that includes the earthquakes generated by stresses, which arise during magma migration in the Earth's crust. They correspond to A- and B-type events of Minakami's classification. These earthquakes may be caused by shear or tensile fracturing during magma movement from depth to the Earth's surface through conduits and dykes. Volcano-tectonic earthquakes occur within and around the volcanic edifice and reflect the interaction of two general geological processes: the magma migration to the Earth's surface and the crustal tectonic activity. In Chapter 3, seismo-generating processes occurring mainly within the flowing magma were discussed; this chapter is devoted to the nature of seismic events occurring during the fracturing of host rocks where magma migrates to the surface.

### 4.1 MIGRATION OF MAGMA AND ITS SEISMIC POTENTIAL

Fig. 4.1 gives a general scheme of magma melting in the mantle and its rising through the ruptured crust to the surface at the zones of ocean spreading and subduction. Magma is formed in the mantle in two principal ways: the first way involves simple decompression of hot, solid mantle material and the second involves lowering the melting temperature of the mantle material by the addition of volatiles (Perfit and Davidson, 2000). Plate tectonics show (Fig. 4.1) that magma is generated mainly below divergent plate boundaries (oceanic spreading ridges) and below convergent plate boundaries (subduction zones). Intraplate volcanism is supported by magma that is generated within so-called *hot spots*-producing *mantle plumes*. The location of the magma generation constrains its physical properties and constitution.

Partial melting of the mantle generates basaltic magmas, and partial melting of crustal rocks generates silicic magmas (Rogers and Hawkesworth, 2000). Common magmas vary nearly continuously in composition from rhyolitic (75% SiO<sub>2</sub>) to basaltic (50% SiO<sub>2</sub>). Generally, three types of magma are distinguished: acid (dacitic–andesitic) containing more than 66% of

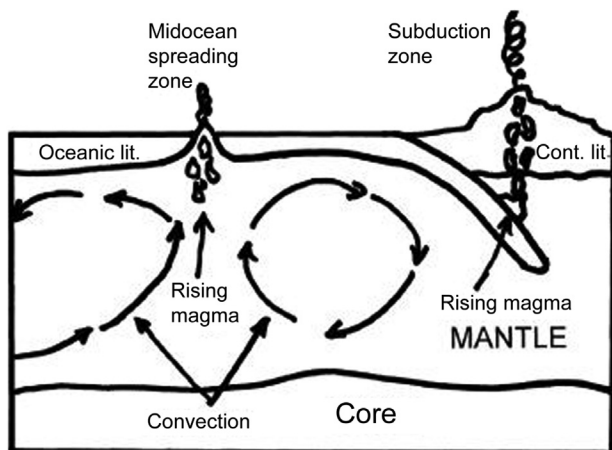


**FIGURE 4.1** Illustration of the process of magma melting in the mantle and its rising at the mid-ocean spreading centers and subduction zones to the surface. Volcanoes overlie the zones where slowly moving oceanic lithosphere is pushed beneath the thicker lithosphere of the continent. The engines that drive this movement are rising heat currents in the mantle. The molten magma rises buoyantly toward the Earth's surface.

$\text{SiO}_2$ , intermediate (andesitic–basaltic) containing between 52% and 66% of  $\text{SiO}_2$ , and basic basaltic containing less than 52% of  $\text{SiO}_2$ .

After magma generation, it commences its journey to the surface. At mantle depths, magma rises very slowly within convection currents without the fracturing of the surrounding material. Within the shallow crust, there is storage of magma, and differentiation begins to occur.

Seismic studies allow the location of the zones of magma storage since they represent a zone of low-velocity and high-attenuation of seismic waves. Figs. 4.2 and 4.3 show the locations of low-velocity zones beneath the volcanoes Pinatubo, Philippines and Mount St. Helens, Cascades. A large zone, representing a relative decrease of  $P$ -wave velocity (5%–10%) beneath Pinatubo volcano (Fig. 4.2), was located between depths of 6 and 11 km and is estimated to have a volume of 40–90  $\text{km}^3$  (Mori et al., 1996). For Mount St. Helens (Fig. 4.3), a  $P$ -wave low-velocity region began at 1.5 km depth, directly below the crater. This localized low-velocity zone extends to a depth of 3.5 km where it then appears to broaden (4–5 km diameter) and continues to a depth of 7.5 km (Lees, 1992). Therefore the regions of magma storage are situated at very shallow depths.

Two types of igneous rocks are distinguished: intrusive igneous rocks and extrusive igneous rocks. Intrusive igneous rocks crystallize when molten rock intrudes into unmelted rock masses deep in the Earth's crust while extrusive igneous rocks are formed when magma erupts at the surface. The intrusive and extrusive rocks can have the same content but different textures. The main types of intrusions are *sills* and *dykes*. Sills are concordant, tabular bodies that are emplaced essentially parallel to the bedding of the country rock and occur

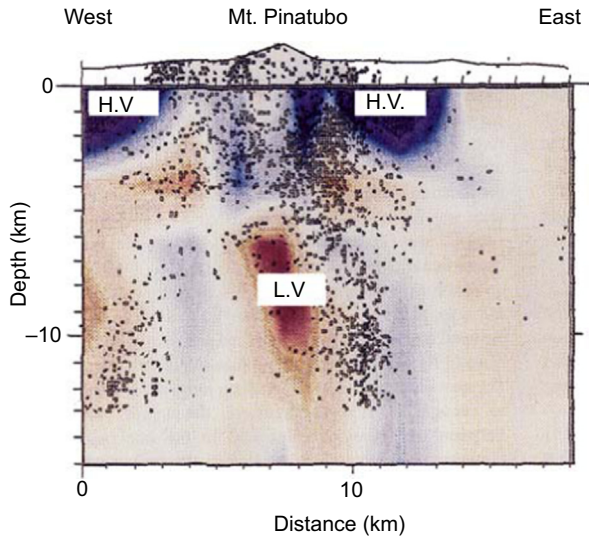


**FIGURE 4.2** Cross-section of P-wave velocity perturbations from the average velocity at a given depth beneath Pinatubo volcano, Philippines. The dots show the hypocenters of posteruption seismicity from June 29, 1991, to August 16, 1991. From Mori, J., Eberhart-Phillips, D., Harlow, D.H., 1996. Three-dimensional velocity structure at Mount Pinatubo: Resolving magma bodies and earthquake hypocenters. In: Newhall, C.G., and Punongbayan, R.S. (Eds.), *Fire and Mud. Eruptions and lahars of Mount Pinatubo, Philippines*. University of Washington Press, Seattle, pp. 371–382.

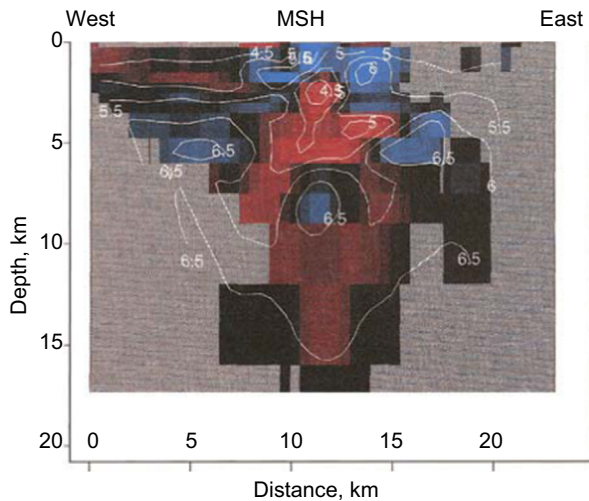


**FIGURE 4.3** Cross-section of P-wave velocity perturbations beneath Mount St. Helens volcano, Cascades. Contours of absolute P-wave velocities (background plus perturbation) are provided for reference. From Lees, J.M., 1992. The magma system of Mount St. Helens: non-linear high-resolution P-wave tomography. *Journal of Volcanology and Geothermal Research* 53, 103–116.



in relatively unfolded regions at shallow crustal levels. A low viscosity of magma is required to produce this sheet-like form. Dykes are thin, tabular, discordant intrusive bodies that cut across the foliation or bedding of the country rock. They range in thickness from less than a meter to several hundred meters, and a few have been traced along strike for tens of kilometers. Typically emplaced into already existing fracture systems, they may occur singly or in swarms (Fig. 4.4). In some areas, dykes occur as radiating swarms centered on an intrusion or on the flanks of a volcano. In rare cases, vertical- or outward-dipping ring dykes or inward-dipping cone sheets are distributed in oval or circular patterns around the intrusion (Blatt and Tracy, 1996).

Extrusive rocks form different types of lavas. The term “lava” is used both for magma that has been erupted onto the Earth’s surface and for the rocks that have been solidified from it (Blatt and Tracy, 1996).

When magma rises from its storage zone, it begins to move along the system of preexisting or newly formed tectonic fractures and the volcanic conduit to the surface as a dyke or sill. The magma ascent rate is a function of the pressure in the magma storage region, the density and viscosity of the magma, the conduit diameter, and the resistance to flow in the conduit that connects the magma storage zone to the surface (Rutherford and Gardner, 2000). All these factors affect the generation of volcano-tectonic seismic activity and define the style of the eruptive process.

Magma is not pure liquid but has rheological properties and its ability to flow depends strongly on its viscosity. Fig. 4.5 shows that the viscosity of different melted rocks at high temperatures (about 1000°C) can change by 10 orders from komatite (less than 1 Pa s) to rhyolite ( $10^{10}$  Pa s; Spera, 2000). This difference in viscosity determines the ability of different magma types to

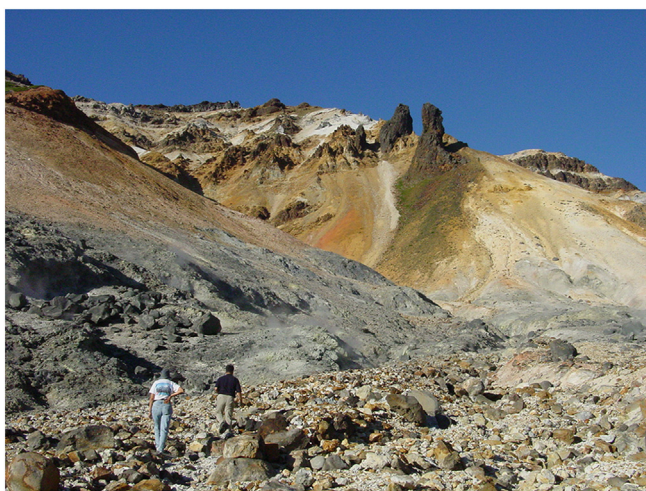


**FIGURE 4.4** Swarm of dykes in the crater area of Tokachi volcano, Hokkaido.



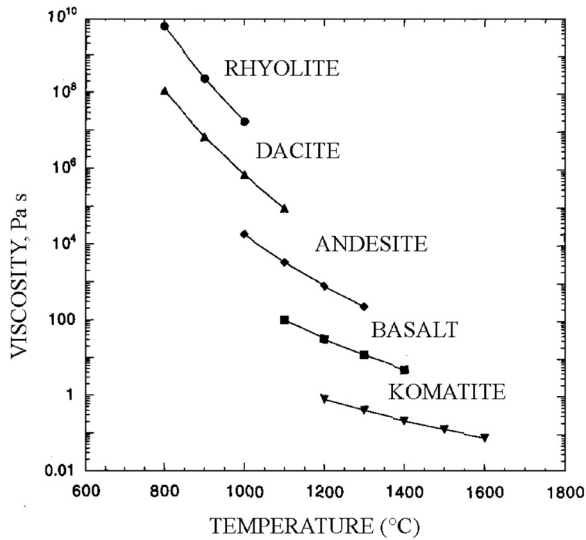


**FIGURE 4.5** Viscosity as a function of temperature at a pressure of 1 bar for natural melts spanning the compositional range rhyolite to komatiite. All compositions are volatile free. From Spera, F.J., 2000. *Physical properties of magma*. In: Sigurdsson, H. (Ed.), *Encyclopedia of Volcanoes*. Academic Press, San Diego, pp. 171–190.

migrate through the cold rocks of the interior of the Earth. Low-viscosity basaltic magma prefers to go along the stratified structures (or faults) in accordance with their strike and does not cut them. The intermediate-viscosity andesite-basaltic magma may choose both ways. High-viscosity dacitic magma slowly cuts stratified structures and forms obelisks. The average ascent rate varies from 0.001 to 0.015 m/s for extrusive magmas. During explosive eruptions, the assessment of magma ascent rate is characterized by much greater variations and leaves the top of the conduit at essentially sonic velocities for truly explosive eruptions (Rutherford and Gardner, 2000). This intrusive–extrusive activity of magma indicates the seismic potential of magmatism within the upper crust.

## 4.2 VOLCANISM AND TECTONICS

Volcanoes normally occur within rift- (or graben) type structures along long-lived faults. A graben is often formed above intruding dykes in volcanic rift zones because a dyke creates two zones of maximum tensional strain at the surface, parallel to the dyke axis (Perfit and Davidson, 2000). The regional tectonic faults where the volcano is situated are accompanied commonly by numerous local faults. These structures could be mid-oceanic ridges or rift valleys (depressions) along the convergent boundary (subduction zone) or above a hot spot (see Fig. 4.1). They could be the basis for a long volcanic

chain or a small volcanic group. Basaltic volcanoes have a well-defined framework of local faults where the flank cones are formed.

Fig. 4.6 shows the structural map of Mount Etna volcano, Sicily (Azarro, 1999). This volcano has grown at the intersection of two regional faults: the Malta Escarpment and the Messina-Fiumefreddo, striking, respectively, NNW and NE, and at the front of the Apennine-Maghrebian thrust belt, the crustal-

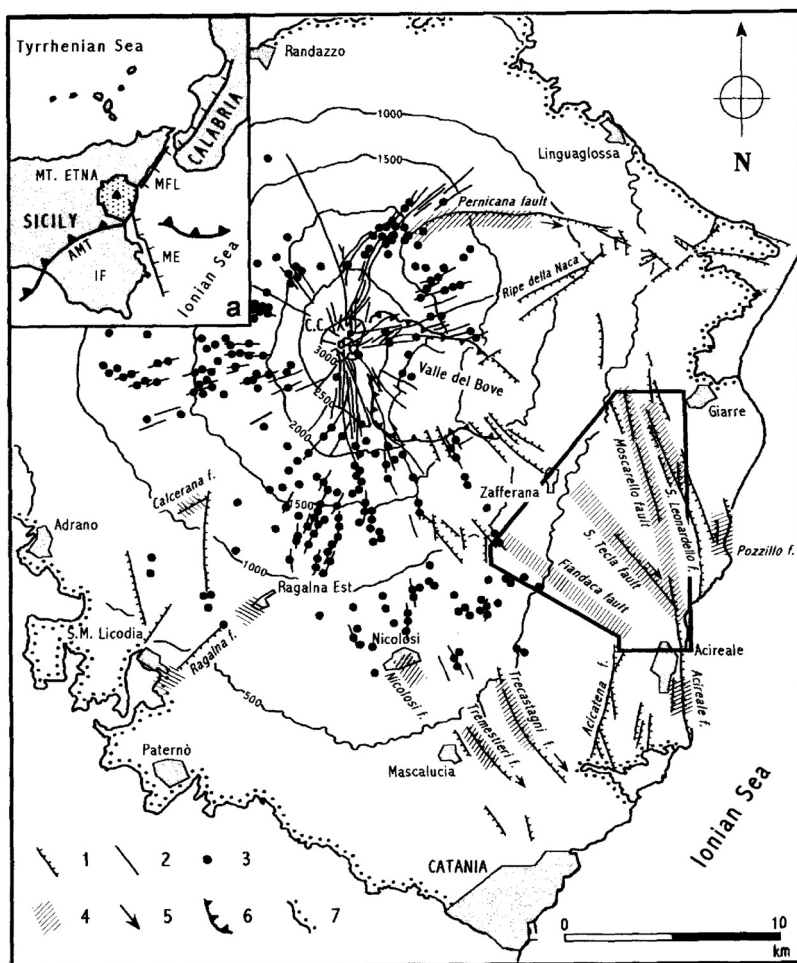


**FIGURE 4.6** Simplified structural map of Mount Etna. (1) Faults with downthrown side, (2) eruptive fissures, (3) cinder cones, (4) coseismic surface faulting zones, (5) strike-slip components, (6) caldera rims, and (7) limit of Etna volcanic edifice. CC, central craters. Box indicates area of the Timpale fault system. Insert map (A) shows the regional geological setting: AMF, front of the Apennine-Maghrebian thrust belt; IF, Iblean Foreland; ME, Malta Escarpment; MFL, Messina-Fiumefreddo line. From Azarro, R., 1999. Earthquake surface faulting at Mount Etna volcano (Sicily) and implications for active tectonics. *Journal of Geodynamics*, 193–213.

scale fault. The tectonic setting of Etna's flanks results from the interaction of regional and local tectonics. The Timple system at the volcano's eastern flank represents the northernmost prolongation of the Malta Escarpment and forms an NNW–SSE-trending system of parallel east-facing step-faults of considerable length (8–10 km). To the north, this system is interrupted by the E–W Pernicana fault that cuts a large part of the volcanic edifice and by the NE–SW Ripe della Naca system. Two isolated structures, Trecastagni and Tremestrieri faults, are situated at the southern flank of the volcano.

The ground breakage that was observed in the volcanic region generally follows preexisting active fault traces. Azarro (1999) also showed evidence of movement along the hidden and semi-hidden local faults. The flank volcanic eruptions occur normally along the local tectonic faults, mainly on the eastern flank of the volcano (Gresta et al., 1990). Among them was the 1991–93 eruption that occurred along the NNW–SSE Timple system of faults (Ferrucci and Patané, 1993). The existence of numerous cinder cones illustrates the process of magma intrusion along the local faults.

Another basaltic volcano, Kilauea, Hawaii (Fig. 4.7) is situated within a complex tectonic framework formed by the Ka’Oki, Koa’E, and Hilina fault

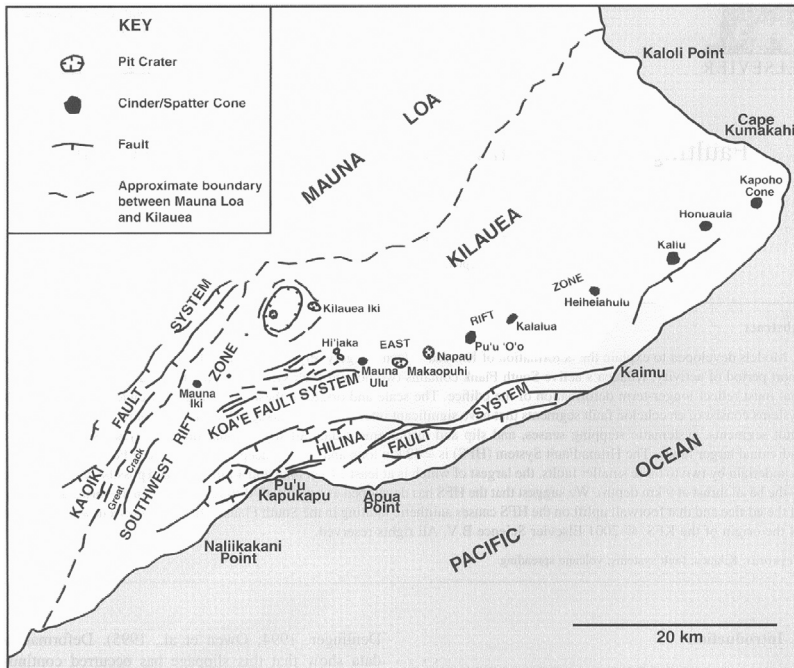


**FIGURE 4.7** Kilauea volcano, Hawaii, and its rifts and fault systems. From Parfitt, E.A., Peacock, D.C.P., 2001. Faulting in the south flank of Kilauea volcano, Hawaii. *Journal of Volcanology and Geothermal Research* 106, 265–284.

systems and the southwest and east rift zones. The normal fault systems at the south flank of Kilauea do not participate directly in the magmatic activity but reflect longer term deformation of the volcanic edifice. The fault systems consist of *en echelon* fault segments that have significant intersections and linkage (Parfitt and Peacock, 2001).

Rift zones, which intersect the flanks of the volcano, are filled by dykes that supply magma and form cinder cones. The dykes propagate downrift at narrow blade-like intrusions from shallow magma reservoirs beneath the summit caldera. Rift eruptions and many summit eruptions begin when the magma issues from long extension cracks that fracture the surface immediately before the eruptions. These eruptive fissures are oriented parallel to the regional trend of the rift zones and mark the intersection of propagating dykes with the Earth's surface (Dieterich, 1988).

The precise location of small (magnitude 1–2) volcano-tectonic earthquakes that were observed during the 1983 dyke intrusion fissure eruption of Kilauea volcano in Hawaii (Rubin et al., 1998) shows a pronounced tightening of seismicity along the dyke trend within the east rift zone (Fig. 4.8A). The foci of earthquakes are located about 0.5 km south of the eruptive fissure, but a dyke that dipped only 7 degrees from vertical would pass through them. The hypocenters of the earthquakes lie between 3 and 4 km depth (Fig. 4.8B) that indicates the origin of seismicity from a very narrow depth interval, even though the dyke breached the surface.

The andesitic stratovolcano Volcán de Colima is located within the N–S-trending Colima Rift zone, in the western part of the Mexican Volcanic Belt, and together with the Pleistocene volcano Nevado de Colima forms the Colima Volcanic Complex (Fig. 4.9). The Colima Volcanic Complex is situated above the subduction zone where the Rivera plate goes beneath the North American plate. At the same time, the Colima Rift represents the eastern boundary of the Jalisco block that is moving toward the ocean from the continent. The Colima volcano marks also the intersection of two large local tectonic structures, the N–S-trending Colima Rift and NE–SW-trending Tamazula fault (Garduño-Monroy et al., 1998).

The November 1998 block-lava extrusion at Volcán de Colima was preceded by a 12-month period of seismic activity that included five volcano-tectonic earthquake swarms (Zobin et al., 2002b). The epicenters of the earthquakes that were recorded during the June–July 1998 swarm clustered across the N–S-trending Colima Rift, while the epicenters of the October–November 1998 earthquake sequence were located along the rift structure (Fig. 4.10). This earthquake clustering may reflect the magma movement along the two local faults that were active at different stages of magma uplift to the surface.

The dacitic volcano Usu, Hokkaido, is situated on the southern rim of the nearly circular Toya caldera that is of about 10 km in diameter, surrounded by a steep wall and partially filled now by Lake Toya. Usu volcano has a characteristic distribution of its central and flank cones (Fig. 4.11) that

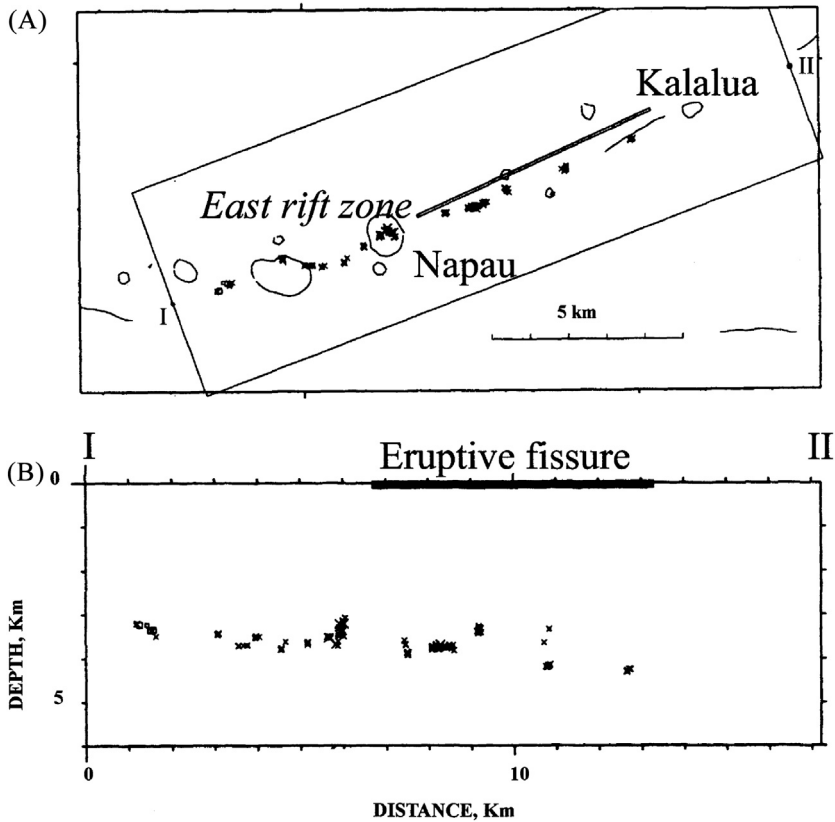


**FIGURE 4.8** Clustering of volcano-tectonic earthquakes. (Crosses) along the eruptive fissure (thick line) during the 1983 dyke intrusion in the east rift zone of Kilauea. (A) Epicenters of earthquakes and (B) cross-section along I–II line of (A). The cinder cones are shown by ovals. From Rubin, A. M., Gillard, D., Got, J.-L., 1998. A reinterpretation of seismicity associated with the January 1983 dike intrusion at Kilauea volcano, Hawaii. *Journal of Geophysical Research* 103, 10003–10015.

outlines a system of circular faults in the volcanic edifice. The craterlets of the 1977 eruptions were formed within the summit caldera, whereas a chain of dacitic cones was formed in 1910 and 1943–44 along the 200-m contour line of the volcanic edifice at a distance of about 2–3 km from the summit (Yokoyama and Seino, 2000). These surface manifestations of volcanic activity reflect the movement of dacitic magma along circular faults.

The earthquakes located during the 1977–78 summit eruption of Usu (Fig. 4.12) show a circular distribution of earthquake clusters around its caldera (Okada et al., 1981). The earthquakes surround an event-free zone beneath the central part of the summit and indicate deep circular fault distribution. Examining a slice at the 0.5 km depth, the epicenters for an arc in the SW part of the volcano are seen, while with the 1.0 and 1.5 km depth slices, the NE arc is outlined.

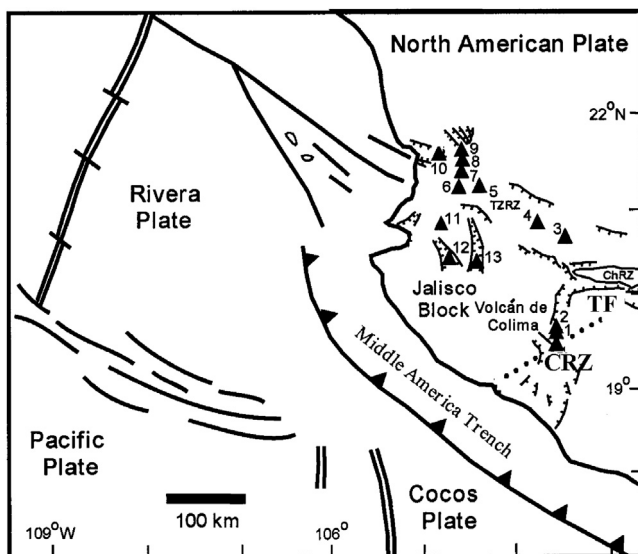


**FIGURE 4.9** Structural position of Volcán de Colima in southwestern Mexico. Three major rift zones are shown: CRZ, Colima Rift zone; ChRZ, Chapala rift zone; and TZRZ, Tepic-Zacoalco rift zone. The dotted line running through Volcán de Colima is the proposed Tamazula Fault zone (Garduño-Monroy et al., 1998). Other major volcanic centers from the western Mexican Volcanic Belt are shown as triangles: (1) Nevado de Colima, (2) Volcán Cántaro, (3) Sierra la Primavera, (4) Volcán Tequila, (5) Volcán Ceboruco, (6) Volcán San Pedro, (7) Volcán Tepetitlic, (8) Volcán Sangangüey, (9) Volcán las Navajas, (10) Volcán San Juan, (11) San Sebastian Volcanic Field, (12) Mascota Volcanic Field, and (13) Los Volcanes Volcanic Field. From Zobin, V., Luhr, J.F., Taran, Y.A., Bretón, M., Cortés, A., De la Cruz-Reyna, S., Domínguez, T., Galindo, I., Gavilanes, J.C., Muñiz, J.J., Navarro, C., Ramírez, J.J., Reyes, G.A., Ursúa, M., Velasco, J., Alatorre E., Santiago, H., 2002. Overview of the 1997–2000 activity of Volcán de Colima, México. *Journal of Volcanology and Geothermal Research* 117, 1–19.

### 4.3 SOURCE NATURE OF VOLCANO-TECTONIC EARTHQUAKES

#### 4.3.1 Waveform and Spectra

Figs. 4.13 and 4.14 show the typical seismograms of volcano-tectonic earthquakes. The high-frequency waveform of volcano-tectonic earthquakes, shown in Fig. 4.13, is characterized by clear arrivals of *P*- and *S*-waves. Both spectra have the dominant frequencies between 5 and 20 Hz but the spectrum of *P*-wave is of higher frequency compared to that of *S*-wave. Low-frequency waveform of volcano-tectonic earthquake (Fig. 4.14) has not so clear arrivals of *P*- and *S*-waves. The dominant spectra of both types of body waves are between 1 and 4 Hz. As usual, low-frequency volcano-tectonic earthquakes are shallower than high-frequency events. Their waveform duration is longer.

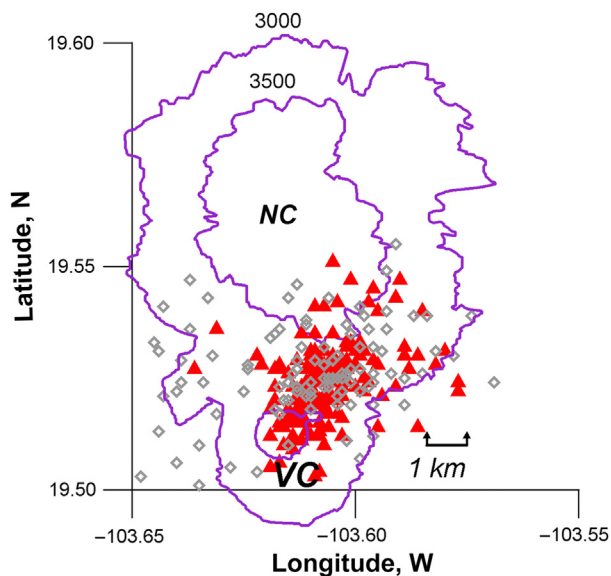


**FIGURE 4.10** Epicenters of earthquakes recorded at Volcán de Colima, Mexico before the November 1998 block-lava extrusion in June 1998 to July (*diamonds*) and October 1998 to November 1998 (*triangles*) NC, Nevado de Colima; VC, Volcán de Colima. The contours of 3000 and 3500 m are shown. *Courtesy of Colima Volcano Observatory*

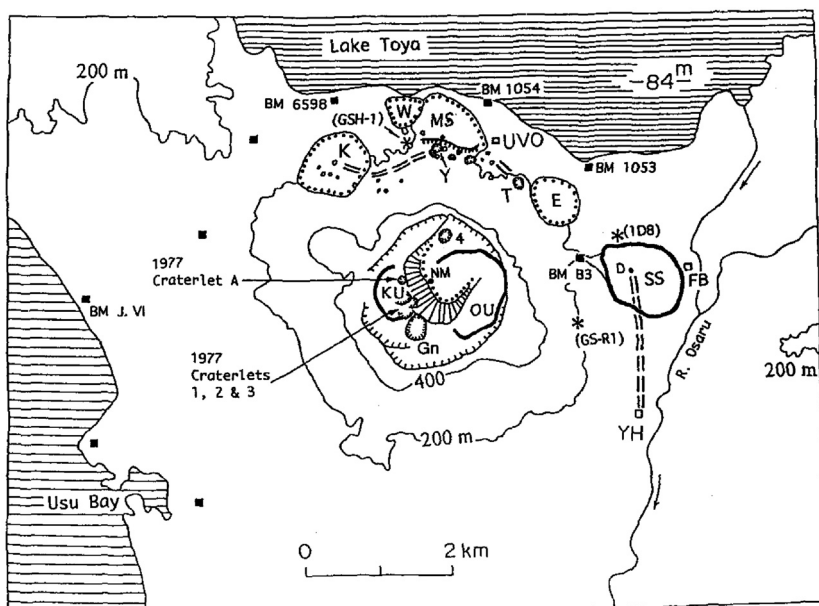


**FIGURE 4.11** Flank cones that were formed at Usu volcano, Hokkaido, during the 1910, 1943–44, and 1977 eruptions. *Contours with dots* show approximate outlines of cryptodomes. MS is the 1910 upheaval; K and E have no dates. *Thick contours* show lava domes. SS indicates the 1943 lava dome; GN and NM are the 1977 Gin-numa craterlet and one of the 1977 peaks, respectively. The stars indicate the wells. *Double broken lines* show the lateral extent of volcanic activity. *From Yokoyama, I., Seino, M., 2000. Geophysical comparison of the three eruptions in the 20th century of Usu volcano, Japan. Earth Planet Space 52, 73–89.*



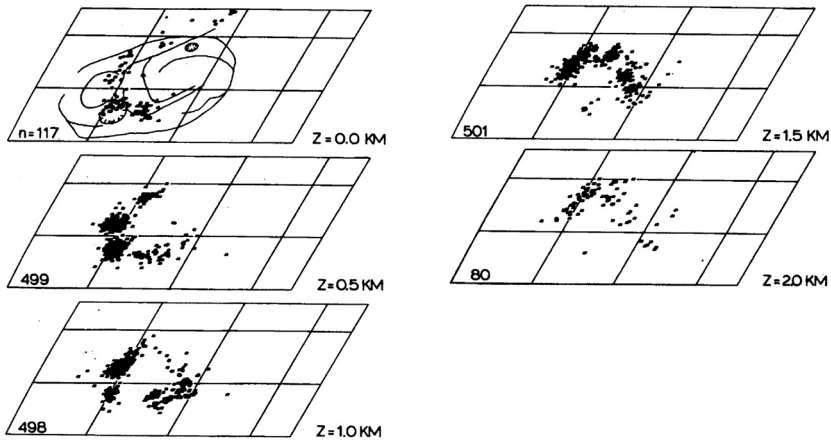


**FIGURE 4.12** Deep slices of earthquake foci located during the 1977–78 eruption of Usu volcano for depths from 0 km (sea level) to 2.0 km. Numbers in the left corner show the number of located events for each depth slice. From Okada, H., Watanabe, H., Yamashita H., Yokoyama, I., 1981. *Seismological significance of the 1977–1978 eruptions and the magma intrusion process of Usu volcano, Hokkaido. Journal of Volcanology and Geothermal Research* 9, 311–334.

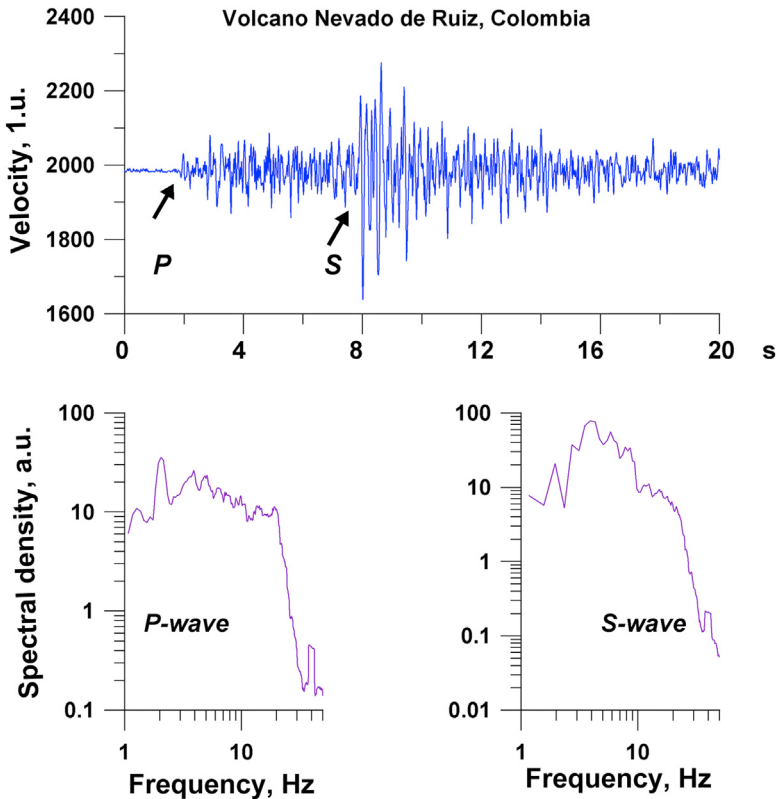


**FIGURE 4.13** Waveform and Fourier spectra of P- and S-waves of high-frequency volcano-tectonic earthquake August 20, 2007, 21:11. Nevado de Ruiz volcano, Colombia.



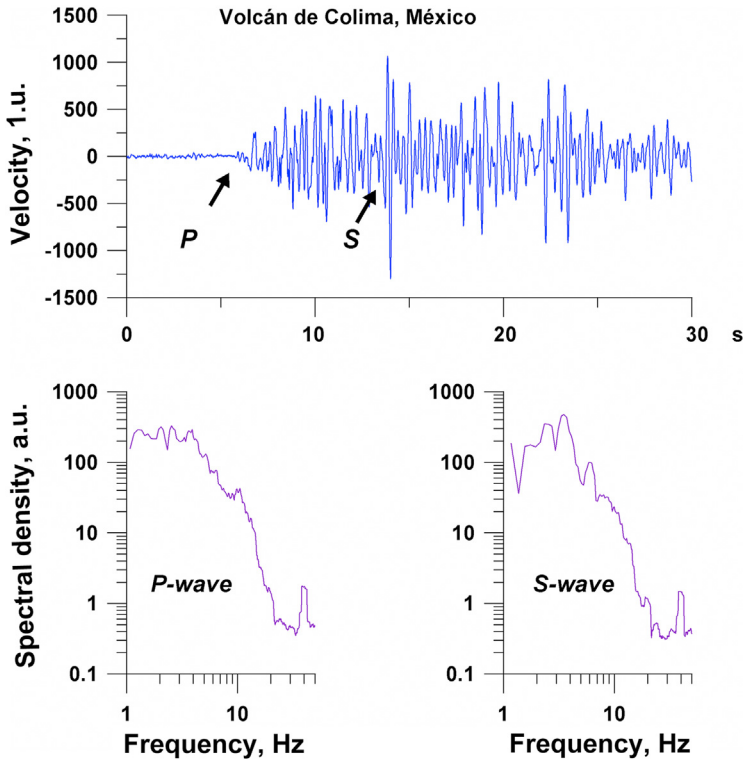


**FIGURE 4.14** Waveform and Fourier spectra of *P*- and *S*-waves of low-frequency volcano-tectonic earthquake November 15, 1998, 10:32. Volcán de Colima, Mexico.

#### 4.3.2 Tensor Representation of the Source of Volcano-Tectonic Earthquake

The complete seismic moment tensor of an earthquake source may be described as a sum of its isotropic and deviatoric parts. The source of volcano-tectonic earthquakes, produced by movement along tectonic fractures, may be represented by its deviatoric part, which is responsible for the production of fracture. The deviatoric part may be decomposed into two source types: a double couple (DC) and a compensated linear vector dipole (CLVD; Fig. 4.15). The dominant type may be discriminated by the relationship between the smallest and largest eigenvalues of the seismic moment tensor  $\epsilon$ ; it is equal to 0 for a DC source and 0.5 for a CLVD source.

Seismological practice and theoretical studies showed the DC source (Fig. 4.16) as the most adequate for shear faulting that produces volcano-tectonic earthquakes. This model predicts the quadrant distribution of *P*-wave first motions using stereographic projection (Fig. 4.17) and allows



**FIGURE 4.15** Model for decomposing the deviatoric part of a general moment tensor. *CLVD*, compensated linear vector dipole; *DC*, double-couple source. Numbers to the left of the diagrams give the principal moments for each mechanism, in units of  $10^{17}$  N m. From Julian, B., Miller, A., Foulger, G. R., 1998. *Non-double-couple earthquakes. 1. Theory. Reviews of Geophysics* 36, 525–549.

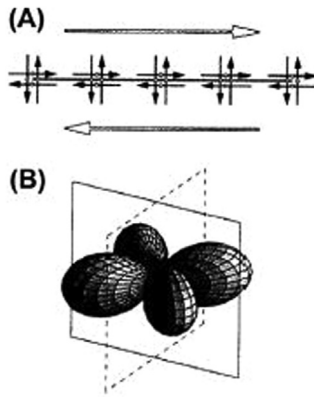


**FIGURE 4.16** Radiation pattern proposed for double-couple earthquake source mechanism. (A) Plane view of a vertical strike-slip shear fault in an isotropic medium, showing the direction of slip (*open arrows*), and the equivalent distribution of double-couple force systems (*solid arrow*, with forces applied at *white dots*). (B) Radiation patterns of compressional *P*-waves, showing fault plane (*solid lines*), and auxiliary plane (*dashed lines*). Adjacent lobes have opposite polarity. From Julian, B., Miller, A., Foulger, G.R., 1998. *Non-double-couple earthquakes. 1. Theory. Reviews of Geophysics* 36, 525–549.

the discrimination of the three main types of faulting in earthquake foci: strike-slip, thrust, or normal fault. The values of  $\varepsilon$  that were obtained for large volcano-tectonic earthquakes related to the eruptions of Mount St. Helens (1980), Myakejima (1983), Oshima (1986), and Teishi Knoll (1989) from the Harvard CMT solutions (Zobin, 1992) varied from 0.10 to 0.18. The focal mechanism solution may be obtained from the direct plots of the *P*-wave first motions on the stereographic projection or by modeling with synthetic seismograms.

The non-DC sources, or the sources with values  $0.2 \leq \varepsilon \leq 0.5$ , were determined initially for a few volcano-tectonic earthquakes located at ring fault structures in Iceland (Ekström, 1994; Nettles and Ekström, 1998) and at Etna and Vesuvius, Italy (Panza and Saraó, 2000; Saraó et al., 2001). The

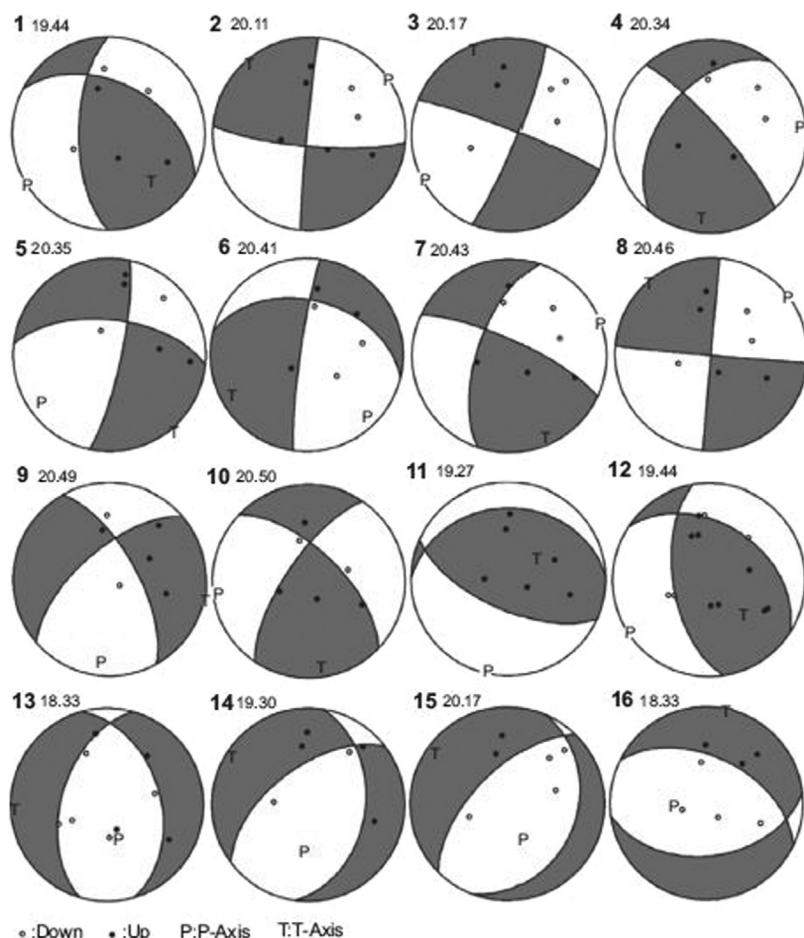


**FIGURE 4.17** Focal mechanism solutions of volcano-tectonic earthquakes from the earthquake swarm preceding the 2000 eruption of Miyakejima volcano. Equal area projection of *P*-wave first motions are projected on the upper hemisphere. Gray and open area show extensional and compressional quadrants, respectively. Mechanisms of strike-slip-type with northeast–southwest stress axis are predominant, although various focal mechanism types are recognized. From Uhira, K., Baba, T., Mori, H., Katayama, H., Hamada, N., 2005. Earthquake swarms preceding the 2000 eruption of Miyakejima volcano, Japan. *Bulletin of Volcanology* 67, 219–230.

focal mechanism solutions for the non-DC sources were obtained by modeling with synthetic seismograms; it has a form shown for CLVD in Fig. 4.15. Minson et al. (2007) found large non-DC components (both isotropic and CLVD) for the 2000 intrusion at Miyakejima and proposed a mixed shearing and opening mechanism of the faults.

Shuler et al. (2013a) had identified 101 earthquakes with vertical CLVD focal mechanisms that took place near active volcanoes all over the world

between 1976 and 2009. As was shown by Shuler et al. (2013b), the physical mechanisms related to fluid flow and volumetric changes are incompatible with seismological, geological, and geodetic observations of vertical CLVD earthquakes.

#### 4.4 MODELS OF VOLCANO-TECTONIC EARTHQUAKE SEQUENCES

As mentioned previously, volcano-tectonic earthquakes occur due to magmatic activity and tectonic fracturing. The movement of magma to the surface occurs within a dyke-type body, and propagating dikes induce migrating seismicity. Rubin and Gillard (1998) proposed a model of dyke-induced seismicity based on dyke intrusion observations at Kilauea volcano, Hawaii (Rubin and Gillard, 1998; Rubin et al., 1998). They study the inelastic outcrop-scale deformation produced within several meters of the dyke tip when the internal pressure variations due to the viscous flow of magma must be considered.

The schematic diagram of the blade-like dyke is shown in Fig. 4.18. In this model, the dyke moves within the host rocks that are subject to the

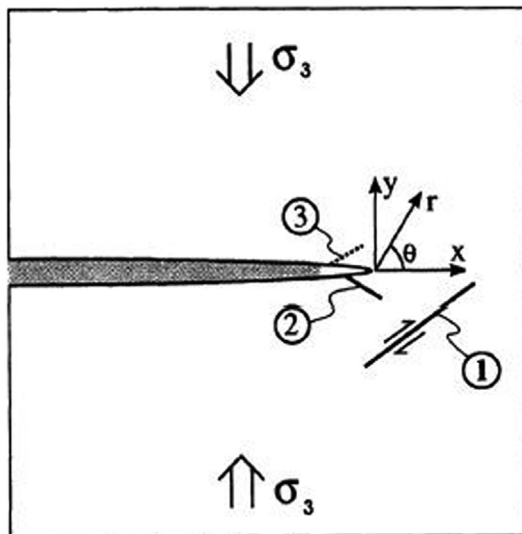


**FIGURE 4.18** Schematic diagram of blade-like dyke (vertical section) and the three types of deformations that could be related to the moving dyke: (1) slip on existing faults away from the tip cavity; (2) slip on existing faults adjacent to the tip cavity; and (3) shear failure of intact rock adjacent to the tip cavity. The least principal stress  $\sigma_3$  is perpendicular to the dyke. The tip cavity of dyke is shown by empty space. From Rubin, A.M., Gillard, D., 1998. Dike-induced earthquakes: theoretical considerations. *Journal of Geophysical Research* 103, 10017–10030.

action of ambient stress. The region near the tip of dyke (tip cavity) where magma cannot penetrate may be occupied by magmatic volatiles or host rock pore fluids. The region influenced by the increase in stress is comparable in size to the length of the tip cavity. The possibility of the magnitude 1–2 earthquakes with a source dimension of about 100 m in the vicinity of the tip cavity is discussed in the following three cases (Fig. 4.18):

- 1 Fault slip away from the tip cavity.
- 2 Fault slip near the tip cavity.
- 3 Shear failure of intact rock.

The examination of elastic stress fields surrounding propagating fluid-filled cracks showed that the stress state is favorable to failure occurring near the tip cavity when the cavity pressure is maintained by the influx of host rock pore fluids, rather than by exsolution of magmatic volatiles. Even in this case, however, shear fracture of previously intact rock seems unlikely. According to these results, most dyke-induced seismicity should result from slip suitably aligned along existing fractures (Rubin and Gillard, 1998).

Rubin and Gillard (1998) suggest that the distribution of earthquake foci of dyke seismicity reflects the distribution of ambient stresses that are near to failure and does not necessarily reflect the extent of the dyke. They reject fracturing as the trajectory of the dyke movement. At the same time, the dyke can trigger the fracturing in the vicinity of its tip along existing fractures. We could consider these fractures as magmatic fractures. Rubin and Gillard (1998) consider that slips could occur at distances over 70–100 m from the dyke if a favorably oriented fault connected two en echelon dyke segments.

The conceptual model of the dyke-induced volcano-tectonic earthquake swarms occurring at basaltic volcanoes was proposed by Hill (1977). It is based on the interaction between dykes and faults. Hill suggested that the clusters of magma-filled dykes exist within brittle volumes of the crust in regions of active magmatic intrusions and the dykes within a cluster are systematically oriented in the regional deviatoric stress field. Their long dimension is parallel to the maximum principal stress  $\sigma_1$ . Hill (1977) considered shear failures occurring along oblique fault planes connecting adjacent tips of offset dykes when a critical combination of the fluid pressure  $P$  in the dykes and the regional stress difference  $\sigma_1 - \sigma_3$  is attained.

Fig. 4.19 shows the schematic representation of the model. The dykes are forced open by a gradual increase in the tectonic stress difference; magma is passively drawn into the dykes. A swarm develops when the stress field within the volume reaches the critical state, and a series of shear failures occur along cracks between the dykes. Hill (1977) suggests that the distribution of earthquake magnitudes will be related to the size and spacing of dykes in the swarm volume in addition to such factors as “roughness” in the variation of frictional properties along individual fractures.

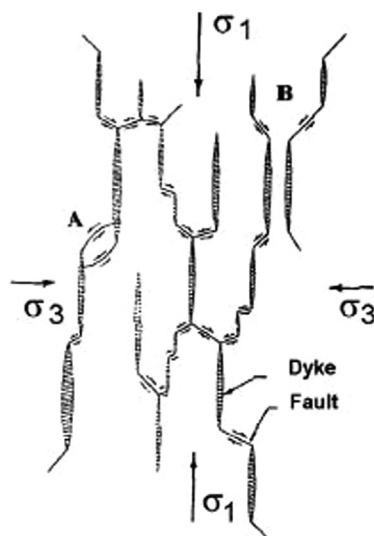


**FIGURE 4.19** Schematic representation of dykes and conjugate fault planes with respect to maximum and minimum principal stresses  $\sigma_1$  and  $\sigma_3$ . (A and B) Typical patterns of crack interactions near adjacent dyke tips in homogeneous media. From Hill, D.P., 1977. A model for earthquake swarms. *Journal of Geophysical Research* 82, 1347–1352.

The next model is based on the stress changes on preexisting faults around the volcanic conduit before an eruption. The stress field rotations obtained from focal mechanism solutions showed that the movement of magma through the shallow crust before an eruption is related to the magnitude and orientation of crustal stresses. [Zobin \(1979\)](#) obtained that during the swarm of volcano-tectonic earthquakes preceding the 1975 Tolbachik fissure eruption in Kamchatka (see Section 5.3.1) this rotation was observed just after two  $M$  4.9 earthquakes occurring in the middle stage of preceding swarm ([Fig. 4.20](#)). [Roman \(2005\)](#) calculated the numerical models of volcano-tectonic earthquake swarms triggering by the magma-filled dykes basing on the inflation-induced Coulomb stress changes on preexisting faults around the magma conduit. Any positive Coulomb stress change was assumed to be sufficient to generate an earthquake. All calculations were made in a homogeneous half-space with elastic moduli appropriate for the shallow crust and an imposed static regional stress field of arbitrary magnitude ( $\sigma_1$  and  $\sigma_3$  horizontal). Dykes were proposed to trend parallel to regional  $\sigma_1$  and inflation was always in the direction of regional  $\sigma_3$ . The modeling of Coulomb stress changes on preexisting faults indicated that a relatively small degree of dyke inflation induces slips on faults in dyke walls with a wide range of orientations, including strike-slip motion with a reversed sense of slip relative to regional faulting.

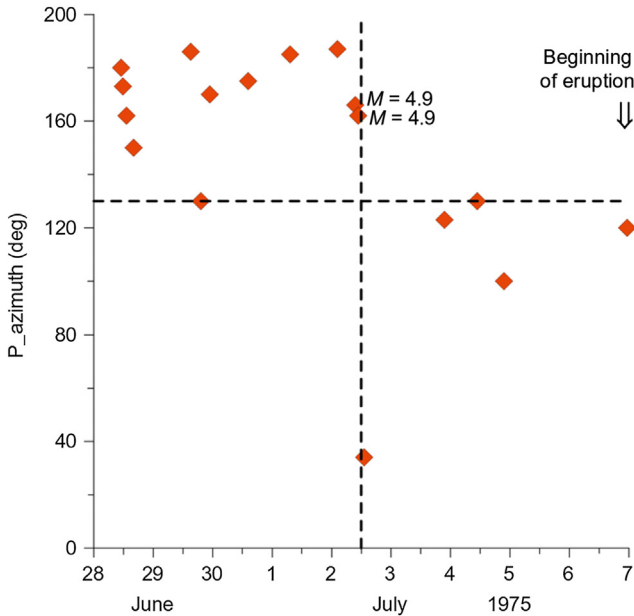


**FIGURE 4.20** Variations in the azimuth of compressional  $P$ -axis obtained from the focal mechanism solutions of volcano-tectonic earthquakes preceding the 1975 fissure Tolbachik, Kamchatka eruption. Dashed lines separate the date before and after the stress field rotation. Modified from Zobin, V.M., 1979. Variations of volcanic earthquake source parameters before volcanic eruptions. *Journal of Volcanology and Geothermal Research* 6, 279–293.

The results of Coulomb stress changes modeling were applied by Roman (2005) to discuss the nature of the “rotated” fault-plane solutions obtained for volcano-tectonic earthquakes occurring during the 1992 Crater Peak (Mount Spurr volcano), Alaska, eruption sequence (Fig. 4.21). The 1992 eruption sequence at Crater Peak was accompanied by volcano-tectonic seismicity over a depth range of 3–10 km b.s.l. Fault-plane solutions for volcano-tectonic earthquakes had  $P$ -axes that were approximately normal to the regional axis of maximum compressive stress (Roman et al., 2004). This local rotation of the stress field was supposed to result from stresses induced by the inflation of a dike-like conduit system.

Coulomb stress changes on the Crater Peak faults were calculated for a series of dike inflation events. The top of the dike was centered directly beneath the vent, the trend of the dike was fixed at N30°W, and all models used a dike height of 10 km. A 1.5 m inflation of a vertical conduit 1 km long and 10 km high (preferred model; Fig. 4.21) resulted in the largest number of Coulomb stress increases on at least one nodal plane out of all models tested. With this preferred model, the majority of the 42 rotated fault-plane solutions experienced a Coulomb stress increase and five earthquakes (14%) experienced large negative Coulomb stress changes ( $> 20$  bar).

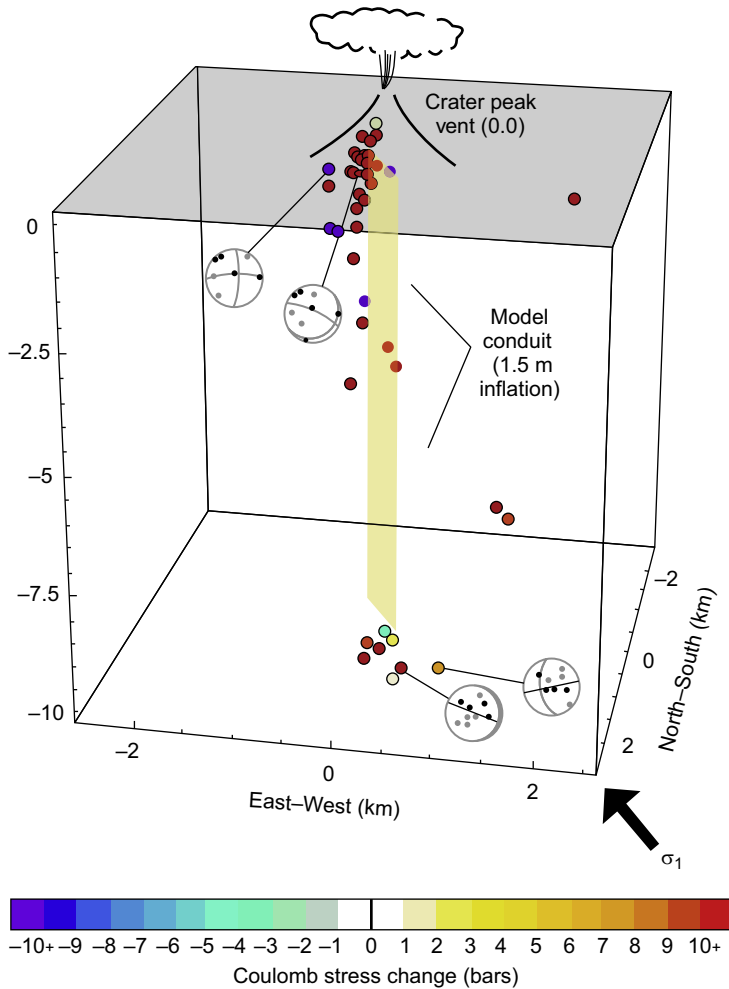


**FIGURE 4.21** Calculated inflation-induced Coulomb stress change on faults represented by “rotated” fault-plane solutions at Crater Peak, Alaska during the 1992 eruption sequence. Drawing shows the locations of earthquakes with rotated fault-plane solutions (dots) relative to the preferred model conduit (described in the text), color coded for the greater calculated Coulomb stress change (on one of the two nodal planes). Lower hemisphere projections of the fault-plane solutions for selected earthquakes (Roman et al., 2004) are shown. Heavy arrow indicates the orientation of regional maximum compressive stress  $\sigma_1$ . From Roman, D. C., 2005. Numerical models of volcano-tectonic earthquake triggering on non-ideally oriented faults. *Geophysical Research Letters* 32, L02304.

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## Chapter 5

# Volcano-tectonic Earthquakes at Basaltic Volcanoes: Case Studies

*Basaltic volcanoes* have characteristic lava of low viscosity ( $10\text{--}10^3$  P), high temperature ( $1000^\circ\text{C}\text{--}1300^\circ\text{C}$ ), with a small amount of volatiles (0.2–2 wt. %), and high velocity of movement. They are characterized by Hawaiian and Strombolian types of eruption and produce extensive lava flows (Vergnolle and Mangan, 2000; Spera, 2000).

The eruptions of basaltic volcanoes may occur from the summit crater of the volcano, but flank or fissure eruptions are very frequent too. The low-viscosity and high-temperature magma are very mobile and able to form the passageway to the surface along or out of the central conduit of a volcano. These properties of magma constrain the possible seismic reaction to the magma movement. The fracturing of rock during the formation of new volcanic vents on the flanks of a volcano or along a new fissure produces volcano-tectonic earthquakes. These earthquakes can be observed also during the movement of magma along the volcanic conduit before the summit eruptions. At the same time, the summit eruptions of basaltic volcanoes may frequently occur without any preliminary volcano-tectonic activity. Lava effusion from the summit and flank (fissure) vents is accompanied by volcanic tremors and explosions.

## 5.1 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH SHIELD VOLCANOES

Shield volcanoes are broad and have gentle slopes. A few shield volcanoes have persistent lava lakes in their summit calderas. Among them are Mount Erebus (Antarctica) and Nyiragongo (Congo). The most active shield volcano is Kīlauea in Hawaii.

### 5.1.1 Kīlauea Volcano, Hawaii

Kīlauea is a tholeiitic shield volcano. Tholeiitic basalt is formed at mid-ocean ridges Kīlauea and oceanic hot spots and makes up much of the oceanic crust. The eruptions of a very liquid magma occur mainly in the summit caldera, the Southwest Rift zone, and the island of Hawaii East Rift zone (ERZ)

(see Fig. 4.7). Formation of vertical dykes appears to be the dominant process in rift-zone intrusion. Kīlauea's volcano-tectonic earthquakes occur beneath the rift zones and summit caldera, at a depth of less than 5 km (Klein and Koyanagi, 1989; Shelly and Thelen, 2019; Shiro et al., 2024).

Volcano-tectonic earthquakes at Kīlauea typically occur in swarms. The history of Kīlauea's volcano-tectonic earthquake swarms, which occurred in 1963–83 (Klein et al., 1987), showed that the swarms can be divided into fast, intense swarms lasting for a few hours to a couple of days (they precede nearly all eruptions of summit caldera and the rift zone intrusions) and slow swarms lasting several days to a few weeks between the eruptive events.

I selected two episodes of volcanic activity: the 1982 short summit caldera eruption and the 2018 Lower East Rift zone (LERZ) eruption and caldera collapse that developed for four months. The catalog of the US Geological Survey (USGS) Hawaiian Volcano Observatory was used for the analysis of the seismic process.

*The 1982 summit caldera eruption* occurred on April 30 and continued for 1 day. The short swarm of earthquakes began about 2.5 hours before the eruption and continued up to the beginning of the volcanic manifestations (Fig. 5.1).

Most earthquakes were recorded during the first hour of activity; then, the number of events gradually decreased. The maximum magnitude event ( $M$  3.2) was recorded just before the beginning of the eruption.

The epicenters of all 47 events recorded between 08:54 hours and 11:41 hours were located within the summit caldera, where the eruptive

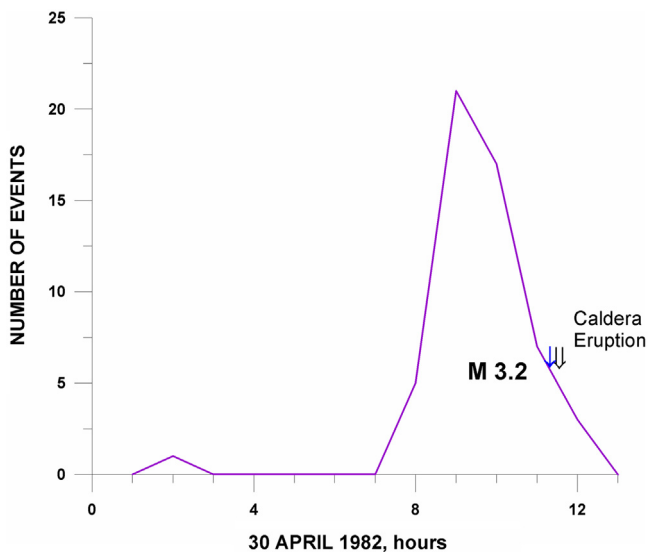


**FIGURE 5.1** Temporal distribution of the earthquakes before the April 30, 1982 Kīlauea caldera eruption. The single arrow indicates the moment of maximum earthquake; the double arrow indicates the beginning of the eruption. Data from the catalog of Hawaiian Volcano Observatory.

fissures opened (Fig. 5.2A). Earthquake depths were estimated within the range from 1 to 2 km beneath the caldera (Fig. 5.2B).

The 2018 Kīlauea eruption began at 2:15 p.m. Hawaii Standard Time (HST = UTC−10) on April 30 (May 1, 00:15 UTC, later we present the date for UTC time) and continued until September 25. The eruption included both a flank fissure eruption and a summit caldera collapse. It was the largest LERZ eruption and caldera collapse in at least 200 years. The 2018 Kīlauea eruption illustrated the association between the summit and rift zones of shield volcanoes. An enormous outpouring of lava occurred in the LERZ and during the concurrent collapse of the summit of Kīlauea situated almost 40 km away. The seismic events, occurring within the swarms of volcano-

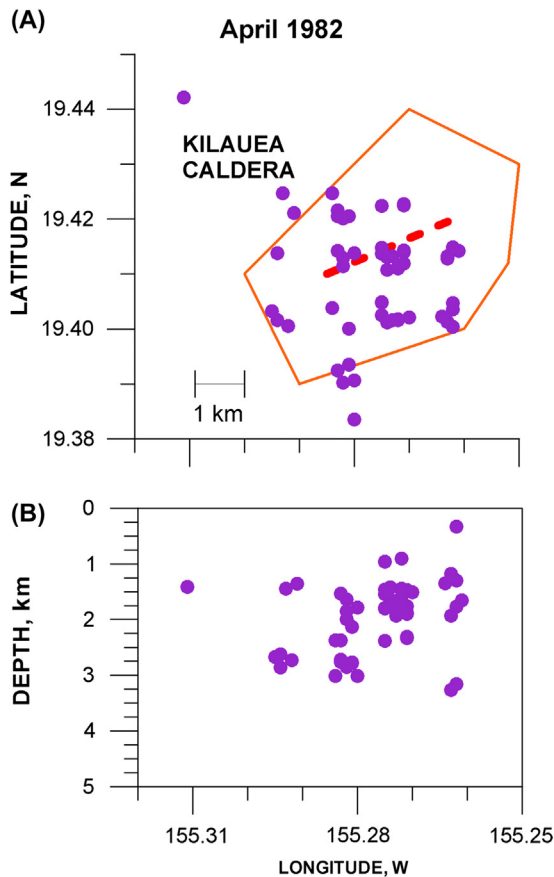


**FIGURE 5.2** Epicenters (A) and hypocenters (B) of the earthquakes recorded on April 30, 1982, at Kīlauea volcano. The crosses show the events recorded between 08:54 and 11:41 hours of April 30 before the caldera eruption. The outline of the caldera (solid lines) and the new eruptive fissures (dashed lines) are shown according to Klein et al. (1987). Data from the catalog of Hawaiian Volcano Observatory.

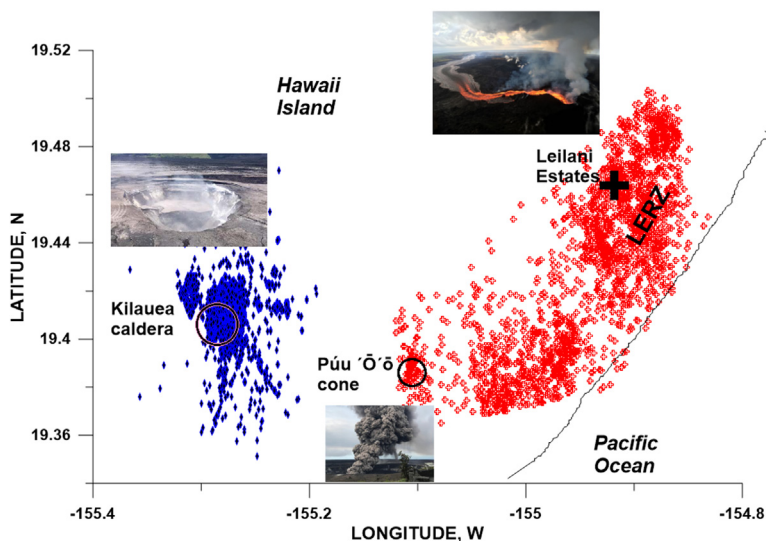


**FIGURE 5.3** Epicenters of volcano-tectonic earthquakes, occurring during the 2018 eruption of Kilauea Volcano from 1 May to 2 August. The events of seismic swarms of Kilauea caldera are shown in blue diamonds, the events of seismic swarms at the fissures of LERZ are shown in red crosses. The positions of the Kilauea caldera and the Pu'u 'Ō'ō crater are indicated with circles. Photos of volcanic episodes represent: a view of the Halema'uma'u crater of the Kilauea caldera (left), the Pu'u 'Ō'ō crater floor collapse (central), and the lava flow from the fissure 8 (right photo). *Courtesy of Hawaiian Volcano Observatory. Seismic data from the catalog of Hawaiian Volcano Observatory.*

tectonic earthquakes, are shown in Fig. 5.3. They were in the areas of Kilauea caldera, Púu Ō'ō cone, and LERZ. The 2018 summit and LERZ volcano-tectonic seismicity was largely concentrated at shallow depth  $<5$  km (Shelly and Thelen, 2019; Shiro et al., 2024) (Fig. 5.3).

The eruption began on May 1 (UTC), with the collapse of the crater floor containing the lava lake in Púu Ō'ō cone, located about 20 km east of the Kilauea caldera and about 20 km west of the LERZ (Fig. 5.3). The collapse of the Púu Ō'ō cone was followed by the subsequent downrift migration of magma toward the LERZ. The development of a seismic swarm, associated with this magma migration from May 1 to May 5, is presented in Fig. 5.4. The map of epicenters of seismic events, occurring from 1 to 3 May (Fig. 5.4A), reflects the directed movement of magma to the village of Leilani Estates, situated in LERZ. As is seen in Fig. 5.4B, during the period from May 1 to May 3, the number of seismic events, reflecting downrift eastward magma movement within a dyke, was high. Practically, a new fissure was formed. The magnitudes of the seismic events reached 2.8 as magma worked its way to the surface to form a new fissure, which erupted on May 3. Subsequently, the number of seismic events decreased, indicating a less restricted path that had been established to allow for

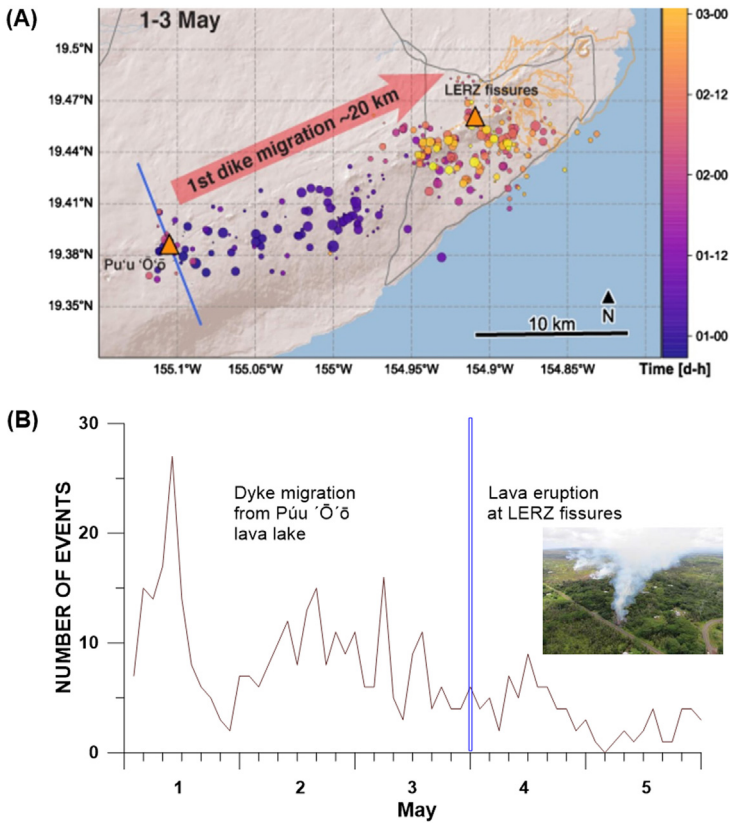


**FIGURE 5.4** (A) Epicenters of the swarm of shallow (depth < 5 km) volcano-tectonic earthquakes originated during the migration of the dike of magma from Pu'u'Ō'ō cone to vents of the LERZ (Modified from Shiro et al., 2024). (B) Temporal development of volcano-tectonic earthquakes of the swarm, occurring during migration of the dike of magma from May 1 to May 5, 2018. In B, the number of seismic events was calculated for 2-hour intervals. The vertical line marks the time of arrival of the dike of magma to the LERZ fissures. (A) The figure is taken from Shiro, B.R., Chang, J.C., Dotray, P., Burgess, M., Thelen, W., Waite, G., Okubo, P.G., Antolik, L., 2024. Earthquake sequences of the 2018 Kilauea Volcano eruption: Observations from the Hawaiian Volcano Observatory earthquake catalog (submitted). (B) The photo of the first fissures erupting in Leilani Estates in B represents a courtesy of Hawaiian Volcano Observatory. Seismic data from the catalog of Hawaiian Volcano Observatory.

the free emission of magma, forming lava flows. Over the next two weeks, beginning on May 3, a series of 24 fissures opened in the LERZ, producing lava fountains and small flows. Lava erupted at rates exceeding 100 cubic meters per second, eventually covering 35.5 square kilometers (Anderson et al., 2024).

After the LERZ eruption was established, the Halema'uma'u crater entered the period of phreatic explosions, which preceded the collapse stage of the eruption. Fig. 5.5 illustrates the development of seismic swarms of volcano-tectonic

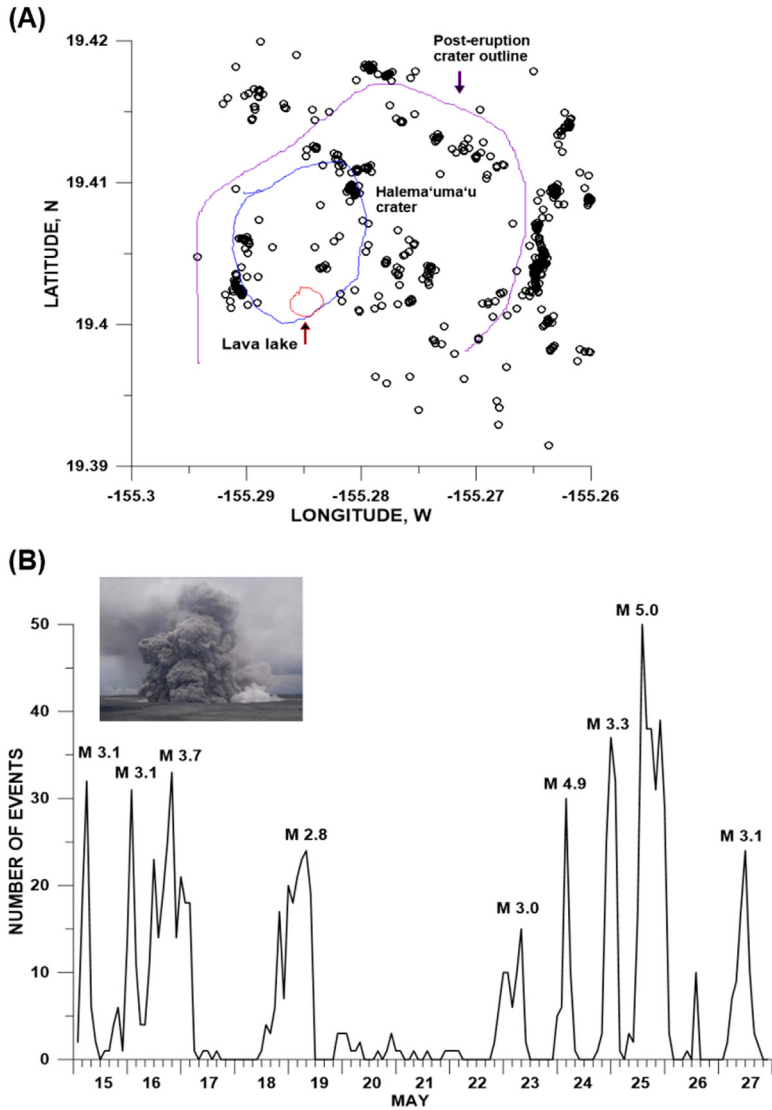


**FIGURE 5.5** (A). Epicenters of the seismic events associated with the phreatic explosive events occurring within the Kīlauea caldera from May 15 to May 27, 2018 during pre-collapse caldera eruption. (B). Temporal development of the swarms of volcano-tectonic earthquakes, occurring within the caldera from May 15 to 27, 2018 and associated with phreatic explosive events. The number of seismic events was calculated for 2-hour intervals. Maximum magnitudes of the earthquake swarms are shown. *Photo of the collapse driven explosion in B represents a courtesy of Hawaiian Volcano Observatory. Seismic data from the catalog of Hawaiian Volcano Observatory.*



earthquakes, associated with this pre-collapse caldera eruption from May 15 to May 29. The seismicity, associated with this activity, was concentrated mainly beneath the structures of Halema'uma'u crater and beneath the eastern portion of the future collapse zone (Fig. 5.5A). Furthermore, nine moments of seismic swarm activity were recorded (Fig. 5.5B). The maximum magnitudes of the seismic events within the swarms varied from 2.8 to 5.0.

The volcano's summit caldera collapse occurred incrementally between the end of May and early August 2018. The overall collapse was not steady but occurred in 62 discrete collapse events, represented by earthquake swarms, with subsidence on the scale of meters. The swarms of volcano-tectonic earthquakes with maximum magnitudes of  $M_w$  5.2–5.4, associated with the collapse caldera eruption, were recorded during the period from May 29 to August 2, 2018 (Butler, 2019). Fig. 5.6A shows the epicenters of these maximum magnitude events. They were situated mainly to the northwest and southeast of Halema'uma'u crater. The strike of the composite earthquake source, derived from stacking events, has a northwest direction. The collapse events punctuated a cycle where hundreds to thousands of smaller earthquakes occurred within the deforming block. The cumulative collapse dropped the crater floor by more than 500 m and enlarged the existing Halema'uma'u crater by over 1500 m in diameter (Butler, 2019; Neal et al., 2019; Segall et al., 2019; Shelly and Thelen, 2019; Shiro et al., 2024).

The temporal development of six seismic sequences, occurring during the initial collapsing of the crater from May 29 to June 7 and characterizing by the appearance of the  $M_w$  5.3 and 5.4 earthquakes, is shown in Fig. 5.6B. The maximum magnitude events were recorded every time as the final event of each of these seismic sequences, occurring at a shallow depth interval from  $-1.1$  to  $0.6$  km (above sea level, a.s.l.). The durations of swarms varied from 8 to 26 hours (for sequences ending in  $M_w$  5.3 events) and from 26 to 42 hours (for sequences ending in  $M_w$  5.4 events) (Butler, 2019).

Caldera collapse and most LERZ lava effusion ended partly on August 2. Summit collapse and lava flow volume estimates were roughly equivalent—about  $0.8$  cubic km (Neal et al., 2019).

## 5.2 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH STRATOVOLCANOES

A stratovolcano is a volcano constructed of alternating layers of lava flows and pyroclastic rocks. Stratovolcanoes are steeper than shield volcanoes and tend to a conical form. The most active basaltic stratovolcanoes are Etna in Sicily, Oshima in Japan, and Klyuchevskoy in Kamchatka, Russia. Many stratovolcanoes have cinder cones and craters of parasitic vents on their flanks (Walker, 2000).

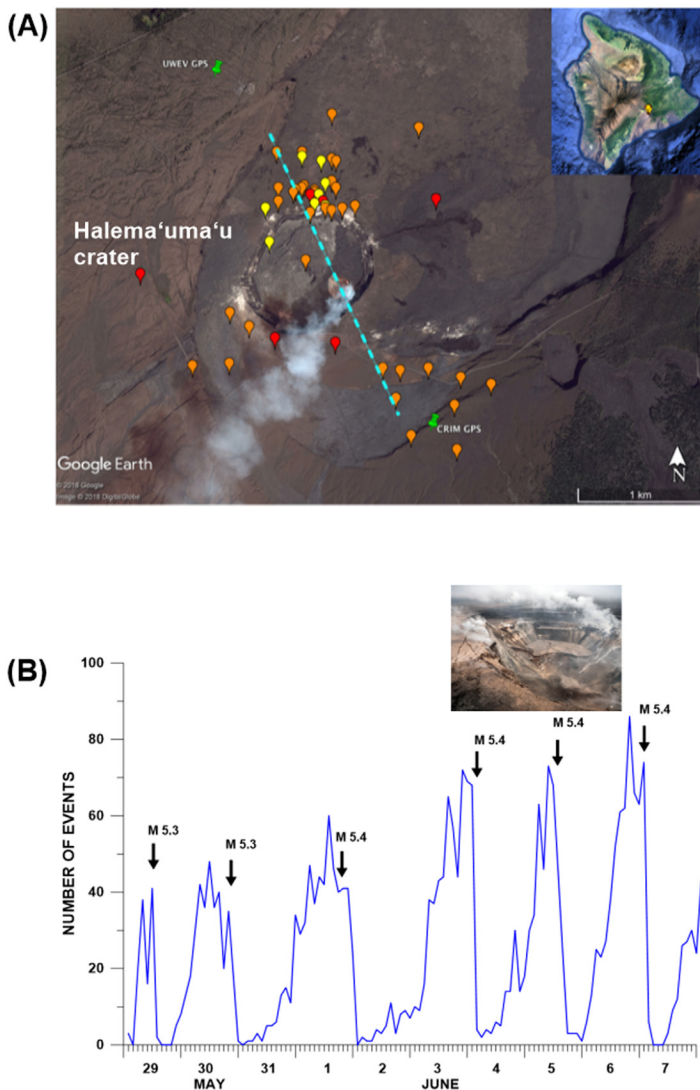


**FIGURE 5.6** (A). Epicenters of seismic events with magnitudes of  $M_w$  5.2–5.4, occurring within the volcano-tectonic earthquake swarms and associated with the collapsing of Halema'uma'u crater within Kīlauea caldera (May 29–August 2, 2018). (B). Temporal development of six swarms of volcano-tectonic earthquakes, occurring within the summit during the initial collapsing of the crater from May 29 to June 7, 2018. In (A), the moment magnitudes  $M_w$  of seismic events are annotated by color (yellow, 5.2; orange, 5.3; red, 5.4). The light blue dashed line approximates the strike of the composite earthquake source derived from stacking events. Inset map locates the Kīlauea events within the island of Hawaii. In (B), the number of seismic events was calculated for 2-hour intervals. Arrows mark the moments of appearance of the  $M_w$  5.3–5.4 earthquakes. The photo shows the enlargement of the collapsed area during June. *Insert map in A modified from Butler, R., 2019. Composite earthquake source mechanism for 2018  $M_w$  5.2–5.4 swarm at Kīlauea Caldera: Antipodal source constraint. Seismological Research Letters, 90(2A), 633–641. <https://doi.org/10.1785/0220180288>. Photo in B represents a courtesy of Hawaiian Volcano Observatory. Seismic data from the catalog of Hawaiian Volcano Observatory.*

### 5.2.1 Mount Etna, Sicily

Mount Etna is a basaltic stratovolcano volcano with an elevation of 3350 m. It is the most active volcano in Europe. Etna activity is characterized by persistent activity at the summit craters (central, NE, and SE craters), as well as episodic flank eruptions (Gasperini et al., 1990). The geodynamic behavior of the eastern and western sides of the volcano is different (Gresta et al., 1990). Earthquake swarms preceded and accompanied the majority of flank eruptions. The summit eruptions were usually not preceded by any seismic crisis (Gresta and Patané, 1987). I selected three episodes of volcano-tectonic activity that preceded the small 1983 lateral eruption and two major lateral eruptions that occurred between 1991 and 2001. I presented also the 1984 summit eruption as an example of the absence of volcano-tectonic earthquakes before the summit eruptions.

*The lateral eruption of March 28, 1983* occurred on the southern flank of the volcano from a short single fracture at the altitude from 2450 to 2250 m (Gresta and Patané, 1987). It continued for 132 days. About  $75 \times 10^6 \text{ m}^3$  of volcanic products (mainly lava) were erupted (Gasperini et al., 1990).

A swarm of volcano-tectonic earthquakes began about 31 hours before the eruption and continued for 28 hours (Patané et al., 1984; Fig. 5.7). It

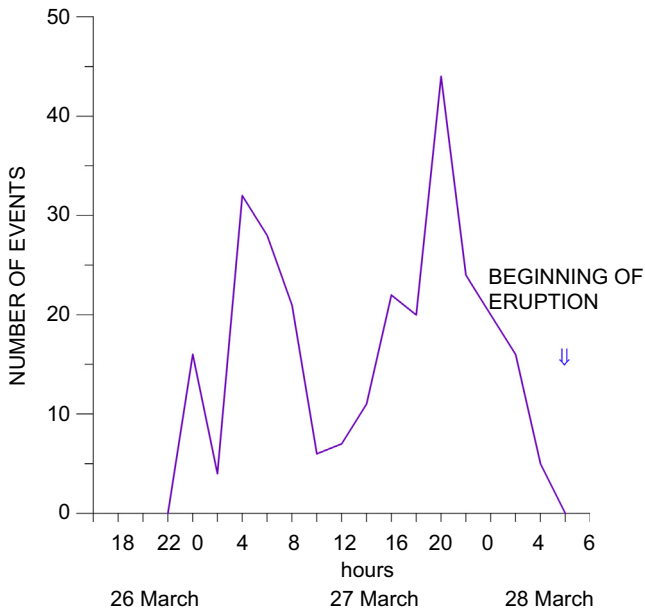


**FIGURE 5.7** Temporal variations of number of the earthquakes before the March 28, 1983 Mount Etna lateral eruption. The number of earthquakes recorded at least by three stations during 2-hour interval is shown. The double arrow shows the beginning of the eruption. From Patané, G., Gresta, S., Imposa, S., 1984. Seismic activity preceding the 1983 eruption of Mt. Etna. *Bulletin of Volcanology* 47, 941–952.

consisted of three impulses. The number of events gradually increased from impulse to impulse during the first 18 hours and reached its maximum about 10 hours before the eruption. Then, the number of events sharply decreased for 6 hours, and the eruption began when seismic activity ceased.

Seismic foci were distributed below the SE slope of the volcano at the depth range from 3 km a.s.l. to 15 km below sea level (b.s.l.) but mainly within the volcanic edifice (Fig. 5.8). The majority of events were concentrated in an area along and on the east of the eruptive fissure (Patané et al., 1984). Patané et al. (1984) consider that only the shallowest events within the volcanic edifice can be related to fracturing processes linked to the opening of eruptive fissures.

*The eruption of April 27, 1984 to October 16, 1984.* “As usual for subterminal eruptions, no seismic crisis preceded the long 1984 SE crater eruption” (Gresta and Patané, 1987). It produced a total lava volume of  $15 \times 10^6 \text{ m}^3$ .

*The lateral eruption of December 1991 to March 1993* was a remarkable event in the eruptive history of Mount Etna, considering both its duration of 473 days and the great volume of emitted products (about  $250 \times 10^6 \text{ m}^3$ ).

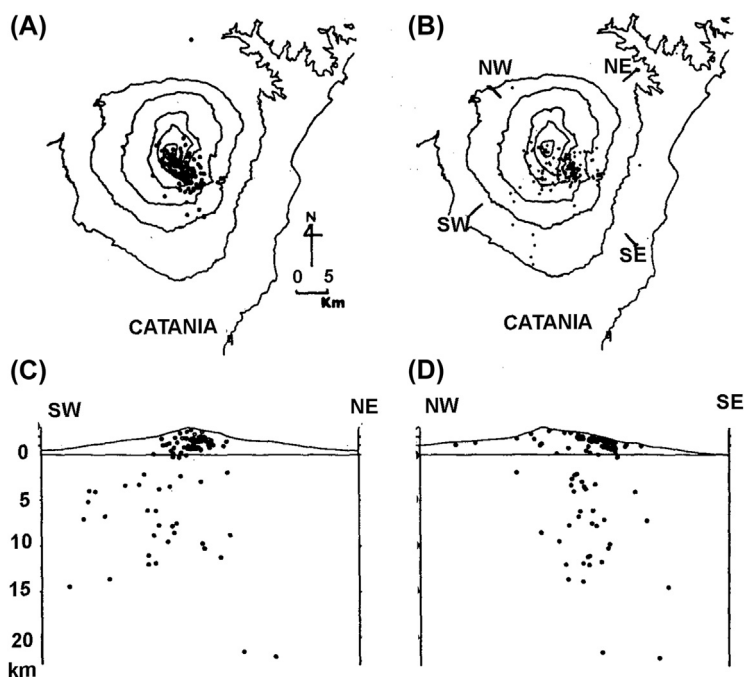


**FIGURE 5.8** Epicenters (A and B) and hypocenters (C and D) of the earthquakes recorded before the March 28, 1983, Mount Etna lateral eruption. (A) The epicentral distribution of the shallowest and smallest earthquakes, (B) the epicentral distribution of the largest earthquakes, and (C and D) the foci distribution along the SW–NE and NW–SE cross-sections. From Cristofolini, R., Gresta, S., Imposa, S., Patané, G., 1988. Feeding mechanism of eruptive activity at Mt. Etna based on seismological and petrological data. In: King, C.-Y., Scarpa, R. (Eds.), *Modeling of Volcanic Processes*. F. Vieweg and Son, Wiesbaden, pp. 73–93.

This eruption took place on the southern flank of the volcano (Lombardo et al., 1996; Fig. 5.9). On December 14, 1991, at about 2:00 hours, the uppermost segment of 7-km-long radial fracture that was formed in 1989 (Ferrucci and Patané, 1993) began to reopen on the flanks of the SE crater, accompanied by a swarm of earthquakes (Figs. 5.9 and 5.10). The majority of earthquakes (about 200 events with magnitude from 1 to 3.3; 17 of them were located by S. Gresta; Fig. 5.9) were recorded during the first 2 hours of

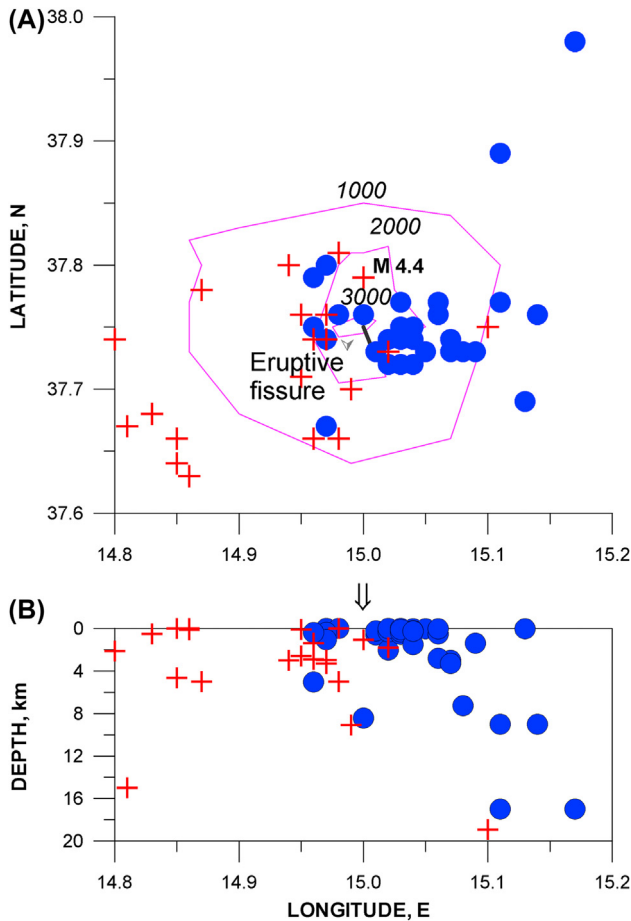


**FIGURE 5.9** Epicenters (A) and hypocenters (B) of the earthquakes associated with the 1991 lateral eruption of Mount Etna. *Filled circles* show the events recorded during the reopening of the 1989 fissures (02:01 to 18:41 hours, December 14); *crosses* show the events recorded during the lava flow from the vents (05:52 hours, December 15 to the end of December). The fissure system on the western slope of the volcano is shown by *heavy line*; its position in (B) is shown by the *double arrow*. The position of the largest earthquake of the swarm ( $M$  4.4; catalog of S. Gresta, personal communication, 2002) is shown, as are the contours of the volcanic relief (1000, 2000, and 3000 m). Data from the catalog prepared by S. Gresta, personal communication.

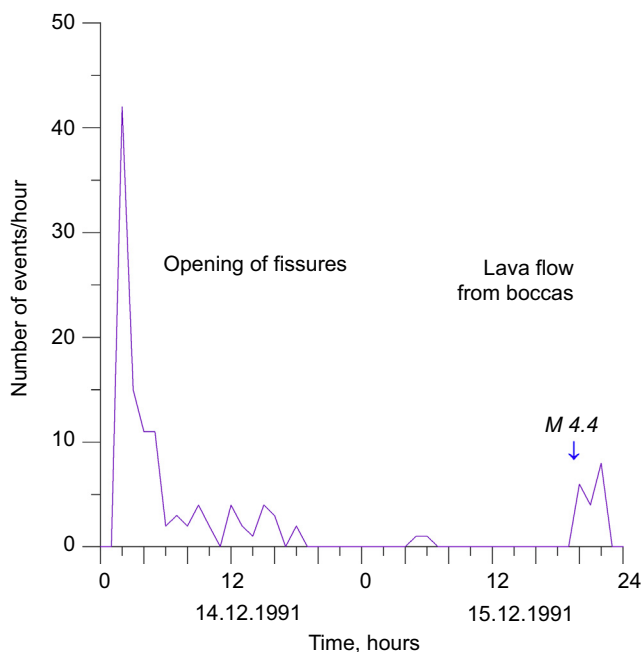


**FIGURE 5.10** Temporal variations of the earthquakes associated with the 1991 lateral eruption of Mount Etna. The arrow shows the onset of the largest earthquake ( $M$  4.4; catalog of S. Gresta) of the swarm. Data from Ferrucci, F., and Patane, D., 1993. *Seismic activity accompanying the outbreak of the 1991–1993 eruption of Mt. Etna (Italy)*. *J. Volcanol. Geotherm. Res.* 57, 125–135.

the propagation of the fissure below the crater and the eastern slope of the volcano at the depth from 0 to 10 km. Then, the seismic activity remained at a low level of 0–3 events per hour.

Early in the morning on December 15, eruptive vents opened at the lower termination of the fracture at the height of 2400 m. A slight increase in seismic activity was observed during the night of December 15 when a sequence of earthquakes occurred below the crater and the western slope of the volcano at a depth of 0–15 km. This hypocentral zone continued its activity up to the end of December. The largest earthquake of the December 14 to December 15 swarm ( $M$  4.4) was recorded during the lava flow from the new vents, 42 hours after the beginning of the swarm.

*The lateral eruption of July 17, 2001* occurred along the 6-km-long submeridional system of fissures that were opened on the southern side of the South-East summit crater at about 3000 m a.s.l. (Fig. 5.11). It lasted for 24 days and produced a total of  $25 \times 10^6 \text{ m}^3$  of lava. In addition, about  $5 \times 10^6 \text{ m}^3$  of tephra were emitted, making this an unusually explosive lateral eruption. The eruption occurred in the background of the sequence of 15 Strombolian bursts occurring from June 7 to July 17 (Alparone et al., 2004; Allard et al., 2006).

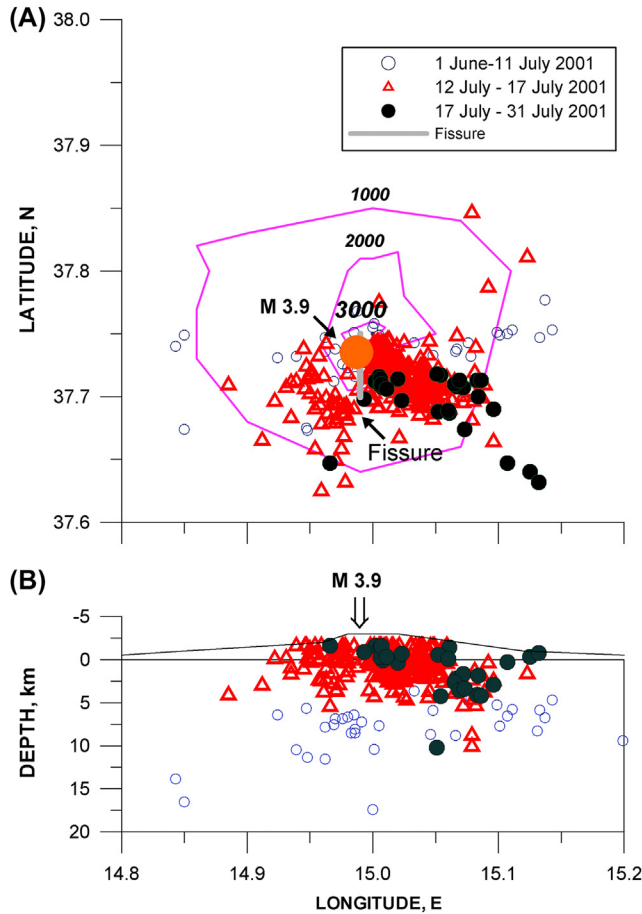


**FIGURE 5.11** Epicenters (A) and hypocenters (B) of the earthquakes associated with the 2001 lateral eruption of Mount Etna. The largest earthquake ( $M$  3.9) epicenter is shown in (A) with the large circle. Its projection is shown also in (B) with the arrow. The fissure zone is shown according to Alparone et al. (2004). Data from the catalog of the National Institute of Geophysics and Volcanology, Section Catania.

Fig. 5.12A shows the temporal variations in volcano-tectonic activity from June 2001 to July 2001 recorded by the seismic network of the National Institute of Geophysics and Volcanology, Section Catania (Patanè et al., 2003). The threshold of the catalog is  $M \geq 1.5$ . The epicenters and hypocenters of these seismic events are shown in Fig. 5.11. It is seen that during the Strombolian burst activity, the daily number of volcano-tectonic earthquakes was between 0 and 7. Seismic activity sharply increased on July 12 reaching 173 events on July 13; one of them was the largest of the seismic swarm events with magnitude  $M$  3.9 (it occurred in 5 hours after the

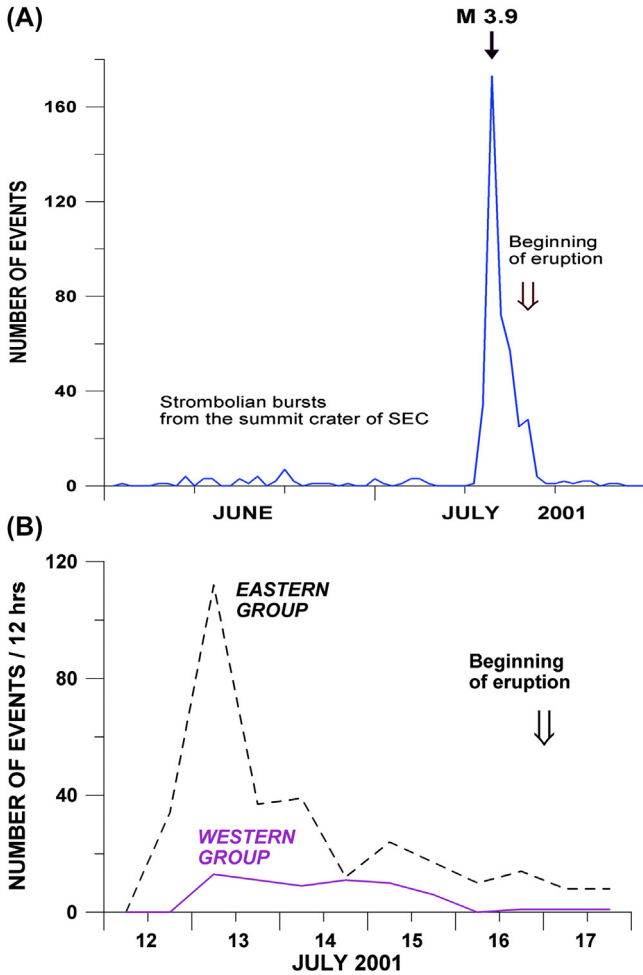


**FIGURE 5.12** Temporal variations of the earthquakes associated with the 2001 lateral eruption of Mount Etna. (A) All earthquakes recorded from June 2001 to July 2001. (B) Earthquakes recorded from July 12, 2001, to July 17, 2001, within the eastern and western epicentral areas. The arrow indicates the onset of the largest earthquake ( $M$  3.9) of the swarm and the double arrow indicates the beginning of the eruption. Data from the catalog of the National Institute of Geophysics and Volcanology, Section Catania.

beginning of the swarm; Fig. 5.12). From July 14 to July 18, the number of earthquakes gradually decreased, and the lateral eruption began on July 17 from six eruptive fissures, accompanied by strong explosive activity and destructive lava flows. After the beginning of the eruption and till the end of the eruption on August 9, daily seismicity stayed at a low level (0–2 events with magnitude  $M \geq 1.5$ ).



Seismicity associated with the background Strombolian activity, recorded from June 7 to July 17, was located along the broad ENE–WSW zone going through the summits of Etna and distributed at the depth range from 5 to 20 km b.s.l. (Fig. 5.11). The preliminary July 12–17 swarms of volcano-tectonic earthquakes clustered into two compact groups of events, western and eastern, separated by a submeridional fissure zone. Fig. 5.12B shows that the activity began on the eastern side but 1 day later the western and eastern groups of earthquakes developed simultaneously. At the same time, the events of the eastern group were more numerous. The foci of both groups of events were distributed at depths from 5 km b.s.l. to 2 km a.s.l., mainly within the volcanic edifice. The largest  $M$  3.9 earthquake was located between these two groups of earthquakes within the fissure zone, at the depth of the sea level. The earthquakes, recorded during the lateral eruption at depths from 5 km b.s.l. to 2 km a.s.l. were located within the eastern cluster and on its continuation in SSE direction. No visible fissures were reported for this direction but maybe they were formed at the depths without opening on the surface.

## 5.2.2 Oshima Volcano, Izu Islands

Oshima volcano is a basaltic stratovolcano that forms an island in Sagami Bay, about 100 km SSW from Tokyo. The eruption of the Oshima volcano started on November 15, 1986, within the central cone of the Oshima caldera (Mihara-yama). A new cone, A, was formed on the wall of the caldera producing a lava flow. On November 21, a fissure opened on the caldera floor at the NE foot of Mihara-yama (vent B); 1.5 h later, a new series of vents (vents C) were opened outside the caldera rim, on the northern slope of the volcano. The activity of new craters continued with some interruptions up to November 18, 1987 (Aramaki, 1998).

Volcano-tectonic earthquakes were recorded during all stages of this 1-year sequence of summit and flank eruptions (Watanabe, 1988; Yamaoka et al., 1988). Fig. 5.13 shows the epicenters of earthquakes located before the summit eruption of November 15 (A), during this eruption (B), and just before the lateral fissure eruption of November 21 (C).

Before the beginning of the summit eruption, the seismic activity of the caldera region was very low and stayed at the same level during the eruption. No preceding swarm was recorded. In contrast, 2 hours before the fissure eruption a swarm of very shallow (depth  $<5$  km) earthquakes began at the northern part of the caldera, and the seismic foci migrated toward the north (Fig. 5.13C). The chain of craters B and C opened along the western part of the swarm area.

A very intensive swarm accompanied fissure eruptions (Figs. 5.14 and 5.15). It occupies a large area of about 40 km<sup>2</sup>. The earthquake foci were distributed at shallow depths (0–5 km) below the island and formed clusters at the depths from 5 to 10 km to the SE (about 10–15 km from the volcano) and to the NW (about 5–10 km from the volcano). Yamaoka et al. (1988)

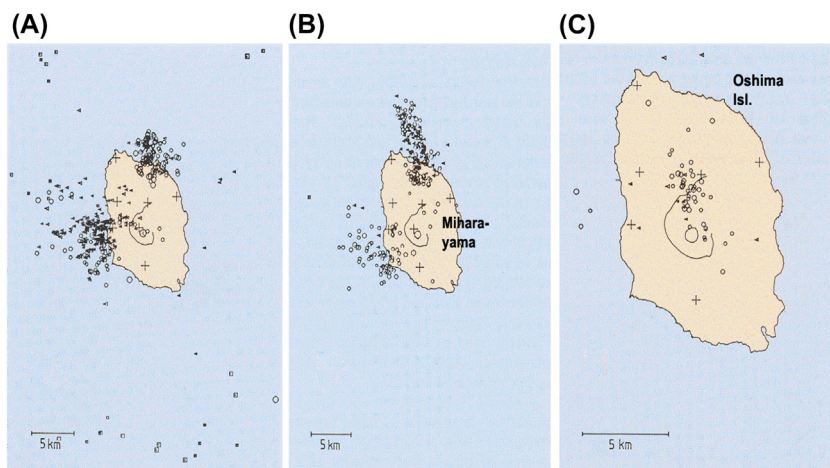


**FIGURE 5.13** Epicenters of the earthquakes located before the November 15, 1986 summit eruption at Oshima volcano (A, from April 20, 1986, to November 15, 1986), during the eruption (B, from November 15 to November 20) and just before the November 21, 1986 fissure eruption (C) of Oshima volcano. From Yamaoka, K., Watanabe, H., and Sakashita, S., 1988. *Seismicity during the 1986 eruption of Izu-Oshima volcano. Bulletin of the Volcanological Society of Japan* 33, S91-S101.

noted that the foci of the events located in these two clusters propagated during the first day of the fissure eruption simultaneously to the SE and NW directions from the eruptive site with the velocity of 11–16 m/min. These two hypocentral zones were active up to the end of March 1987. The largest earthquake of the swarm ( $M_w$  5.9) occurred on November 22, about 18 hours after the beginning of fissure eruptions, off the SE coast of the island.

### 5.2.3 Klyuchevskoy Volcano, Kamchatka

Klyuchevskoy stratovolcano is the most active basaltic volcano of the Kurile-Kamchatka volcanic zone. It is the highest active volcano in Eurasia with an altitude of 4750 m. Its activity is represented by paroxysmal and small summit eruptions and frequent flank eruptions. The summit eruptions of Klyuchevskoy occur with the formation of cinder cones within the summit crater followed by explosions and lava flows.

The paroxysmal eruption of 1945 was preceded by significant seismic activity; the shocks were felt at a distance of 32 km from the volcano a few hours before the eruption of January 1 (Piip, 1956). The small summit eruptions that occurred during 1949–86 were not preceded by seismic activity that could be recorded at a distance of 16 km where the nearest seismic

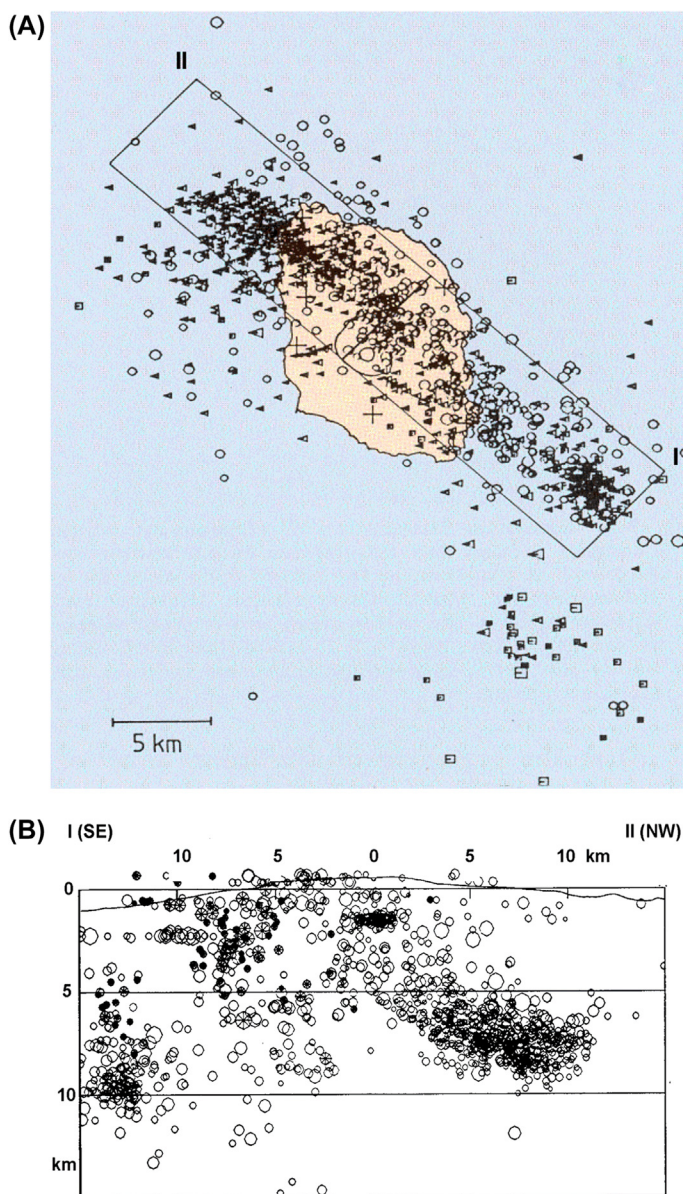


**FIGURE 5.14** Seismic activity accompanied the fissure eruption of Oshima volcano from 16:15 hours on November 21, 1986, to November 30, 1986. (A) Epicenters of earthquakes and (B) hypocenters of earthquakes. I–II is the zone of the cross-section of B. From Yamaoka, K., Watanabe, H., and Sakashita, S., 1988. Seismicity during the 1986 eruption of Izu-Oshima volcano. *Bulletin of the Volcanological Society of Japan* 33, S91-S101.

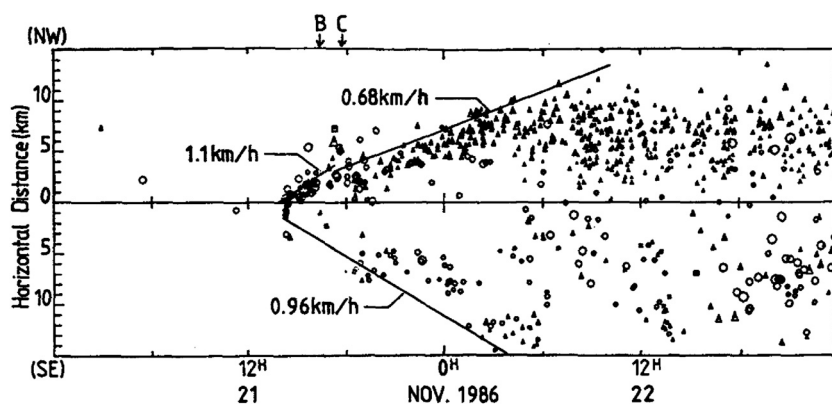


**FIGURE 5.15** Temporal variations of epicenters from November 21 to November 22 in the region shown in Fig. 5.14A. The focal depths are shown by circles (0–5 km), triangles (5–10 km), and squares (> 10 km). The difference in magnitude is shown by the difference in the size of the symbol. Arrows B and C denote the time of the B and C craters' eruptions. From Yamaoka, K., Watanabe, H., and Sakashita, S., 1988. *Seismicity during the 1986 eruption of Izu-Oshima volcano. Bulletin of the Volcanological Society of Japan* 33, S91-S101.

station was installed; only volcanic tremor was observed before and during the summit eruptions (Tokarev, 1966; Khrenov et al., 1991).

The 12 lateral (flank) eruptions that occurred on the flanks of Klyuchevskoy from 1932 to 1987, at altitudes from 450 to 3900 m (Khrenov et al., 1991) were preceded by seismic activity. We discuss here the seismic activity associated with two flank eruptions, of 1951 and 1966.

*The November 19, 1951 Bylinkina flank eruption* occurred at the height of 950 m with the formation of two new cones followed by lava flow (Tokarev, 1981). Seismic activity began 6 days before the beginning of the eruption (Fig. 5.16). The number of earthquakes increased rapidly during the first 30 hours, and then gradually decreased for three and a half days. The strongest earthquake with a magnitude of 3.1 occurred 2 days after the beginning of seismic activity. The eruption began 1 day after the ceasing of the seismic activity. Only one station KLY was in operation, so no locations were made.

*The October 6, 1966 Piip flank eruption* occurred at the altitude of 1900–2100 m with the formation of a sequence of explosion craters (Tokarev et al., 1968). The seismic swarm began 4 days before the eruption with three magnitude 3.5–3.7 events located at depths of about 20 km during the first hour of activity. All remaining events were very shallow, with their magnitudes ranging between 0 and 3. In total, 131 earthquakes were located. The epicentral zone was divided into two parts, the SE deep zone and the NW shallow zone, with the site of eruption situated between them (Fig. 5.17).

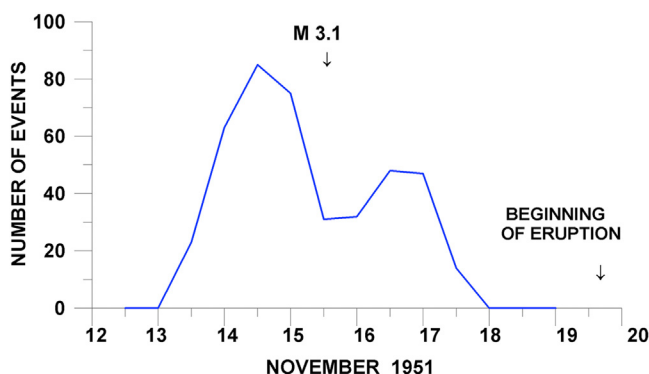


**FIGURE 5.16** Temporal distribution of 12-h number of the earthquakes. ( $M > 1$ ) before the November 19, 1951 eruption of the Bylinkina flank cone of Klyuchevskoy volcano, recorded at the seismic station KLY (see Fig. 5.18), at a distance of 22 km. The moment of the largest event with magnitude  $M$  3.1 is shown. From Tokarev, P.I., 1981. *Volcanic Earthquakes of Kamchatka*. Nauka, Moscow (in Russian). 164 p.

Two peaks in the seismic activity were recorded during the first 3 days of the swarm (Fig. 5.18). Then, the number of events decreased rapidly. The eruption occurred in the background of the low-level seismic activity.

## 5.3 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH FISSURE ERUPTIONS

A fissure eruption is an eruption that takes place from an elongated fissure, rather than from a summit vent (Schmidt and Schmincke, 2000). The most famous fissure eruptions were of Parícutin in Mexico (1943–52), Krafla in Iceland (1975–83), and New Tolbachik in Kamchatka (1975–76). Well-documented episodes associated with dyke injections along rift zones occurred during 2005–09 in the southernmost Red Sea rift near the Afar triple junction.

### 5.3.1 New Tolbachik Volcanoes, Kamchatka

A chain of new volcanoes was formed from July 6, 1975, to December 10, 1976, along the 12-km-long fissure that was opened at a distance of about 18 km SW from Plosky Tolbachik volcano (Fig. 5.19). The new eight volcanoes were basaltic cinder cones with a height of up to 300 m;  $1.2 \text{ km}^3$  of lava was produced by the cones, which covered an area of about  $45 \text{ km}^2$ . The total volume of pyroclastic material was about  $1 \text{ km}^3$ .

The preceding swarm of earthquakes (73 events with magnitude  $M > 2$ ) began on June 27. It is possible to discriminate two stages in the seismic activity (Fig. 5.20): from June 27 to June 30 and from July 1 to July 5.

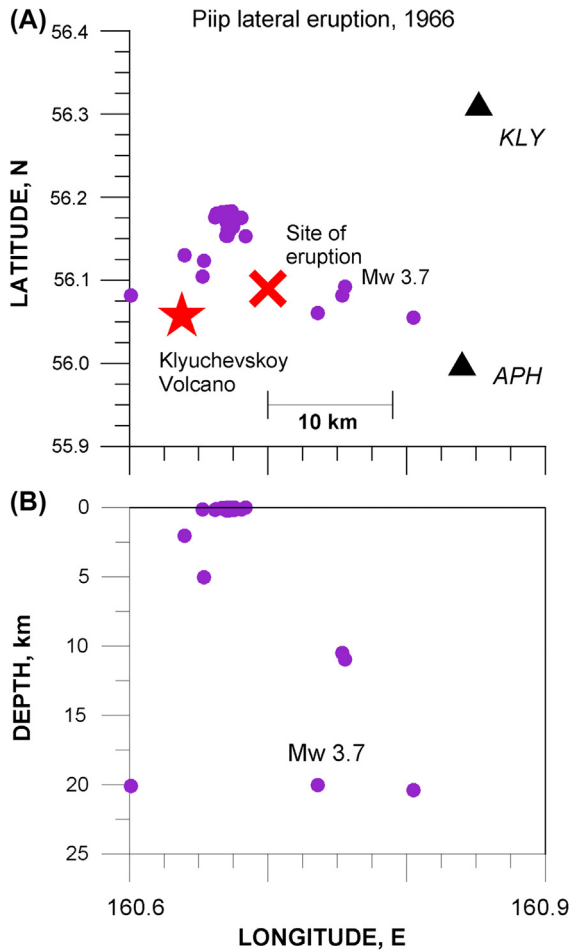


**FIGURE 5.17** Epicenters (A) and hypocenters (B) of earthquakes. ( $M > 2$ ) located before the October 5, 1966 eruption of the Piip flank cone of Klyuchevskoy volcano. Star shows Klyuchevskoy volcano; cross shows the site of the new eruptive center. Two nearest seismic stations are shown by triangles. Data from the Kamchatka Regional Earthquake Catalog.

During the first stage, a sharp increase in the number of events was observed on June 27, and then the number of events gradually decreased for 3 days. The epicenters were distributed within an ellipse-shaped area just to the north of the site of the forthcoming eruption of cone I (Fig. 5.19A) and covered the space where the following six cones appeared. The depths of earthquakes varied from 15 km to the surface (Fig. 5.19B).

The second impulse in seismic activity was characterized by a lower number of events and ceased 1 day before the July 6 eruption. The epicenters occupied



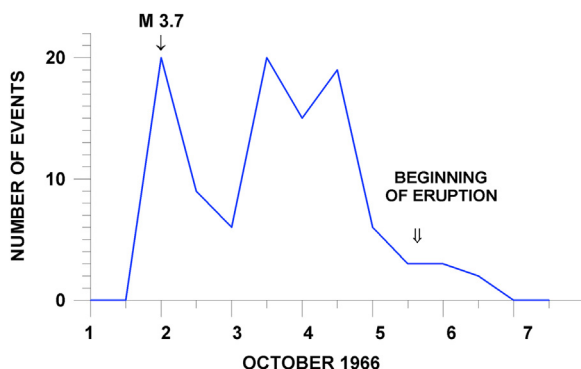


**FIGURE 5.18** Temporal distribution of 12-h number of earthquakes. ( $M > 1$ ) before the October 6, 1966 eruption of the Piip flank cone of Klyuchevskoy volcano recorded at the seismic station KLY (Fig. 5.17) at a distance of 25 km. The moment of the largest event with magnitude  $M$  3.7 is shown. Data from Tokarev, P.I., 1981. *Volcanic Earthquakes of Kamchatka*. Nauka, Moscow (in Russian). 164 p.

a more compact area than during the first stage, nearer to the forthcoming eruption site. They were shallower and were distributed within the depth range from 8 km to the surface. The two largest earthquakes of the swarm (magnitude  $M_w$  4.9) were observed on July 2 at a depth of 11 km. Their epicenters were located below cones II and III. During the following eruptive activity with the formation of the new cones in August 1975–September 1975, a few seismic sequences were observed (Fig. 5.21). One sequence (11 events with  $M > 2$ ) occurred from August 6 to August 8 and preceded the birth of cone II on August 9. The epicenters were distributed within the southern part of the epicentral zone of the first swarm of June 26–July 6, on the level of the forthcoming site of cone II. The next sequence of small earthquakes (35 events with  $M > 2$ ) occurred from August 10 to August 14, before forming the third and the fourth cones on August 17. These seismic events connected the new eruptive zone with the old basaltic Plosky Tolbachik volcano (Fig. 5.21). The caldera collapse, occurring at Plosky Tolbachik in the middle of August (Dvigalo et al., 1984), marked clear relation between the old volcano and the new active cones.

The last seismic swarm (12 events with magnitude  $M > 2$ ) was recorded from September 16, 1975, to September 18, 1975, preceding the birth of the cone VIII (Fig. 5.21). The location of these events continued the active line from Plosky Tolbachik volcano through the North Zone of cones I–VII (they finished their activity on September 15) to the south up to the site of cone VIII, which was born on September 18.

The total process of migration of epicenters along the 30-km line Plosky Tolbachik–Cone VIII from August 6 to September 18 is seen in Fig. 5.22. During the initial part of the post-I-cone seismicity (August 6–8), the epicenters were clustered on the position of the forthcoming cone II. Then, after

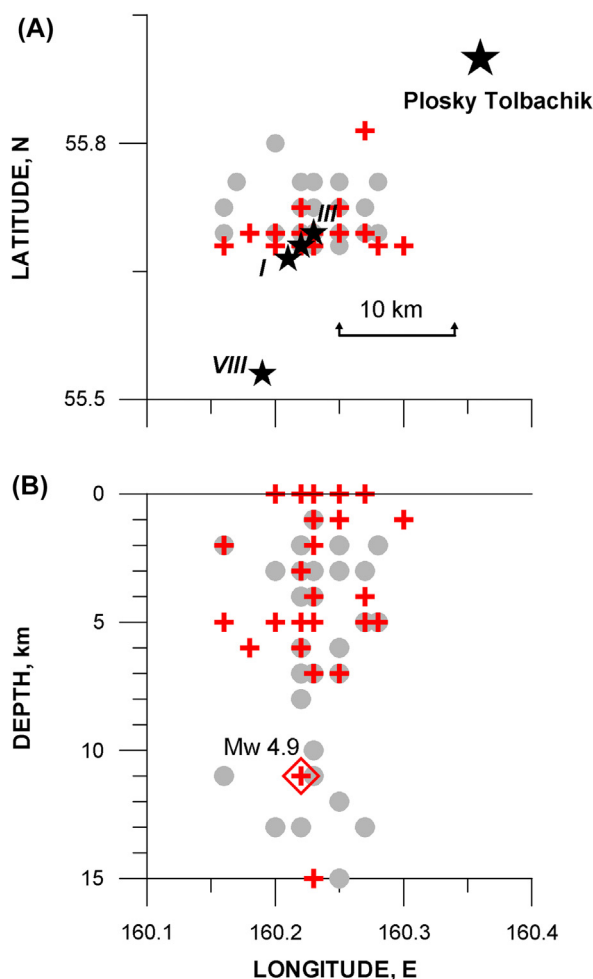


**FIGURE 5.19** Epicenters (A) and hypocenters (B) of the earthquakes. ( $M > 2$ ) preceding the July 6, 1975 eruption of the first cone of New Tolbachik volcanoes. The epicenters of the events recorded from June 26 to June 30 are shown by filled circles; the events recorded from July 1 to July 6 are shown by crosses. The largest earthquake is shown by diamond. Active volcano Plosky Tolbachik (PT in B) is shown by large star; new cones I, II, III, and VIII are shown by stars. Data from the Kamchatka Regional Earthquake Catalog.

forming cone II on August 9, an NE migration of foci was observed to the Plosky Tolbachik volcano, which continued from August 10 to August 14 and coincided with the caldera collapse of Plosky Tolbachik. The last SW focal movement was from the site of cones I–VII to the site of the forthcoming cone VIII. After the end of activity of the eighth cone on December 15, 1976, seismic events began to spread along the line of new volcanoes, to the



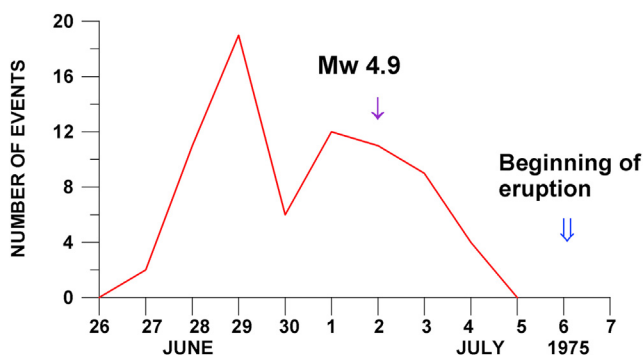


**FIGURE 5.20** Temporal variations of daily number of earthquakes. ( $M > 2$ ) preceding the July 6, 1975 eruption of the first cone of New Tolbachik volcanoes. The moment of the largest event is shown. Data from the Kamchatka Regional Earthquake Catalog.

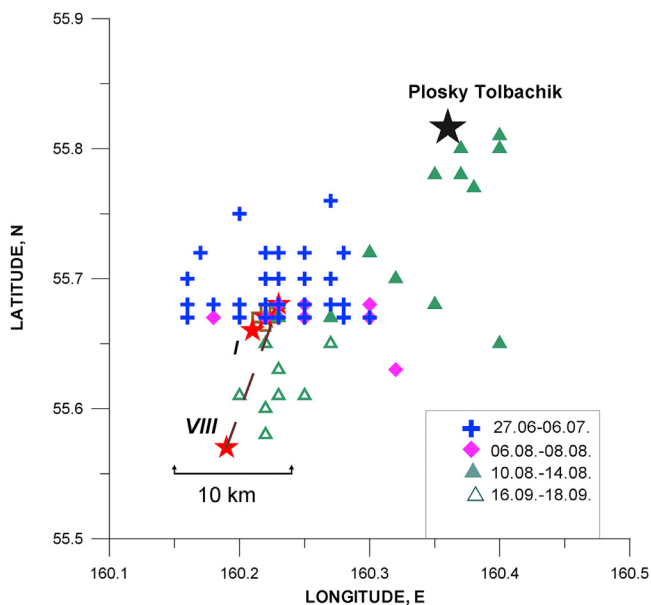


**FIGURE 5.21** Epicenters of earthquakes. ( $M > 2$ ) that were observed during the June 1975 to September 1975 eruption of the New Tolbachik cones. Four groups of events are shown: June 26 to July 6 (crosses); August 6–8 (diamonds); August 10–14 (filled triangles), and September 16–18 (open triangles). Cones I–VIII are shown by small stars; active volcano Plosky Tolbachik is shown by large star. The fissure zone is shown by dashed line. Data from the Kamchatka Regional Earthquake Catalog.

north and south from the active centers (Fig. 5.23). During 2 weeks, a line of 90-km length was outlined by the epicenters.

Posteruption seismic activity was recorded during 1977 and 1978 at the periphery of the active volcanic zone at a distance of 40–60 km from the new

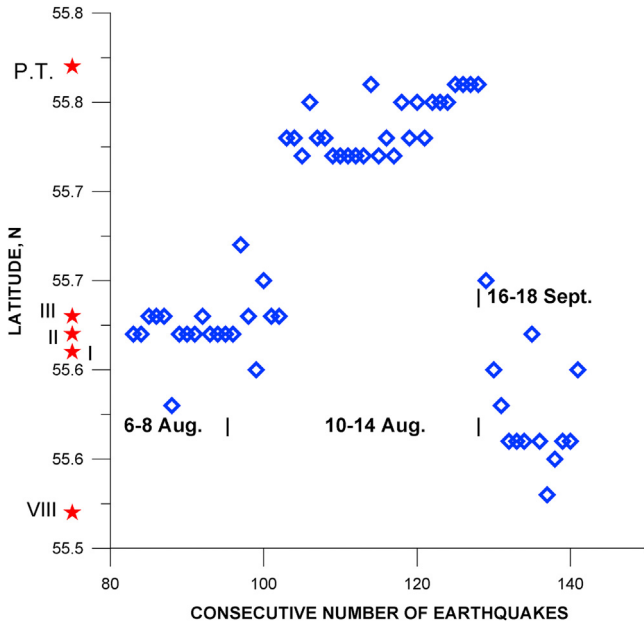


**FIGURE 5.22** Migration of earthquake foci. ( $M > 2$ ) along the NE line Plosky Tolbachik (PT) volcano—Cone VIII from August 6 to September 18 Stars along the latitude axis show positions of the new cones (I, II, III, VIII) and PT volcano. Data from the Kamchatka Regional Earthquake Catalog.

volcanoes. The magnitudes of the largest posteruption earthquakes reached 5.2. This seismic activity is supposed to be caused by the discharge of hydraulically connected magma chambers that fed the eruption (Zobin, 1990).

### 5.3.2 The 2005–09 Ethiopia Rifting Episode

The 2005 Dabbahu, Ethiopia rifting episode occurred in the southernmost Red Sea rift near the Afar triple junction. Afar region was affected by the seismotectonic crisis in 1978 and 1989 (Ayele et al., 2007). Dabbahu is a Holocene volcanic massif forming an axial range of the Afar depression. The obsidian flows, lava domes, and pumice cones form the summit and upper flanks of the volcano, which was built over a base of basaltic-to-trachytic lava flows of a shield volcano. Late-stage Holocene basaltic fissure eruptions occurred at the NW base of the volcano. Abundant fumaroles are located along the crest of the volcano and extend NE. The first historical eruption of the Dabbahu volcano took place from September 26, 2005, to September 30, 2005, from a fissure vent on the NE flank of the volcano. It was a small explosive eruption that produced ashfall deposits. The 2005 event ruptured the length of a magmatic segment within the Dabbahu-Manda-Hararo segment of the Manda-Hararo rift

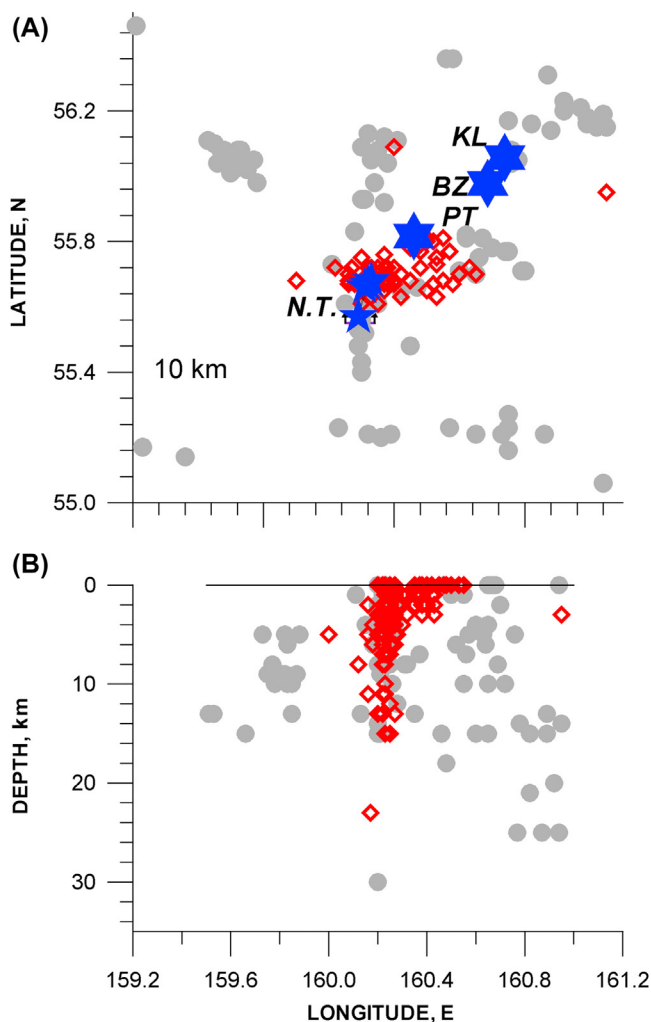


**FIGURE 5.23** Posteruption (from December 16, 1976, till 1978) seismic activity in the region of New Tolbachik volcanoes. (A) Epicentral distribution of earthquakes and (B) hypocentral distribution of earthquakes. *Open diamonds* are earthquakes recorded during the active period (July 27, 1975, to December 15, 1976); *filled circles* are earthquakes recorded from December 16, 1976, to the end of 1978. The *five-ray stars* show New Tolbachik volcanoes; active volcanoes of the Klyuchevskaya group (KL, Klyuchevskoy; BZ, Bezymianny; and PT, Plosky Tolbachik) are shown by *six-ray stars*. NT is the site of the New Tolbachik volcanoes. Data from the Kamchatka Regional Earthquake Catalog.

zone. It was preceded by at least two preliminary earthquake swarms recorded in 1978 and 1989 (Ayele et al., 2007; Ebinger et al., 2010).

The seismic crisis began on September 4, 2005 (Ayele et al., 2009). Activity subsided but then restarted on September 20 and continued until October 4

(Ebinger et al., 2010). The 2005 epicentral zone was developed between the Ado'Ale volcanic complex (AVC) and the Dabbahu-Gab'ho volcanic complex (DVC; Fig. 5.24). Two impulses of seismic activity were observed: from September 20 to September 23 and from September 23 to September 26 (Figs. 5.24 and 5.25). Nine earthquakes with magnitudes  $M_w$  between 5.0 and 5.5 were recorded. The largest  $M_w$  5.5 event occurred on September 24 at the peak activity of the second seismic impulse. The September 26 eruption began when the number of earthquakes decreased (Fig. 5.25). The September 2005 eruption vent was situated approximately 5 km ENE of the Dabbahu volcano summit (Fig. 5.24; Ayele et al., 2007).

Subsequent fissure basalt intrusions, displayed by continuous earthquake swarms (Fig. 5.24), were recorded from June 2006 to June 2009 along the Dabbahu-Manda-Hararo rift segment, in its central and southern parts

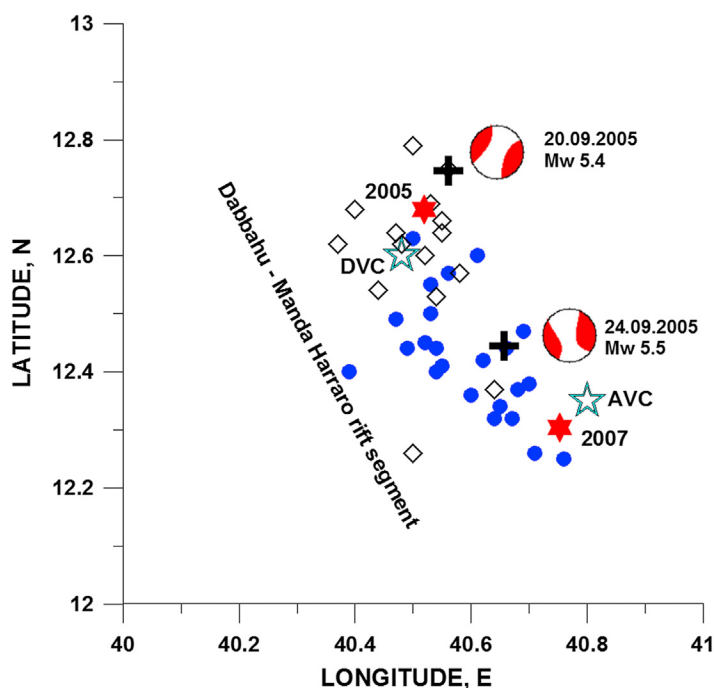


**FIGURE 5.24** Epicenters of earthquakes in the Dabbahu-Manda-Hararo segment from September 20, 2005, to October 2, 2005. The epicenters of earthquakes are shown with *open diamonds* (recorded from September 20 to 23) and *filled circles* (recorded from September 24 to October 2). The Dabbahu (DVC) and Ado'Ale (AVC) volcanic complexes are shown with *open stars*. *Filled stars* show the sites of the September 26, 2005 and the August 14, 2007 eruptions. The epicenters of the largest  $M_w$  5.4 September 20, 2005 and the  $M_w$  5.5 September 24, 2005 earthquakes are shown by *crosses*. Their Harvard CMT focal mechanisms are shown. *Data from the catalog published in Ayele, A., Keir, D., Ebinger, C.J., Wright, T.J., Stuart, G.W., Buck, R., Jacques, E., Ogubazghi, G., Sholan, J., 2009. The September 2005 mega-dike emplacement in the Manda-Hararo (Afar) nascent oceanic rift. Geophysical Research Letters 36, L20306.*

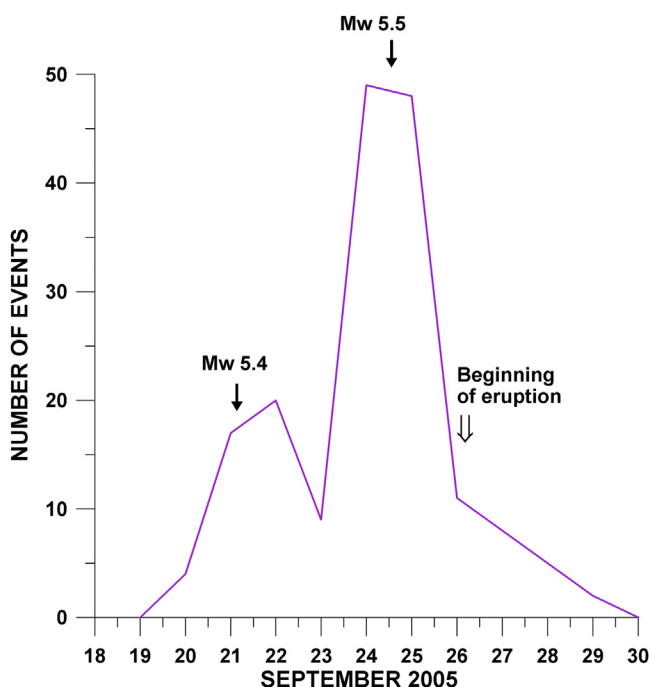


**FIGURE 5.25** Temporal distribution of earthquakes before the September 26, 2005 eruption during the Dabbahu rifting episode. A *single arrows* indicate the moments of maximum earthquakes; the *double arrow* indicates the beginning of magma intrusion. Data from the catalog of USGS.

(Ebinger et al., 2010). The site of the August 14, 2007, mafic eruption in the Karbahi graben south of AVC (Ayele et al., 2009) is shown in Fig. 5.24.

Along the 120-km-long rift segment, 420 from the 2005 to 2009 earthquake hypocenters clustered between 2 and 12 km. The largest of them shows a rift normal opening (Ayele et al., 2009; see also Fig. 5.24). The postdyking earthquakes mimic the geodetically determined distribution of the 2005 dyke (Ebinger et al., 2010).

## 5.4 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH CALDERA COLLAPSE

A caldera is a gigantic basin with steep walls at the summit of a volcano and is larger than a crater. It is usually formed by collapse (Decker and Decker, 1999). The multiple stages of collapse following the evacuation of the underlying magma chamber are characteristics of Kīlauea (Hawaii) and Fernandina (Galapagos Islands). The great 1883 eruption of Krakatoa

(Indonesia) and the 1912 eruption of Katmai (Alaska) are the best geologically documented examples of caldera-forming events.

#### 5.4.1 Fernandina Volcano, Galápagos Islands

Fernandina is a large basaltic shield volcano on the western edge of the Galápagos platform (Fig. 5.26). At its summit, some 1500 m a.s.l., an elliptical caldera of  $4 \times 6.5 \text{ km}^2$  of area is situated. On May 21, 1968, the eruption started, and a lava flow was observed on the east flank of Fernandina. A sequence of strong explosions began on June 8 and then declined on June 12. The aerial reconnaissance of June 21 revealed that the 8-km<sup>2</sup> caldera floor subsided in a series of short drops along an elliptical boundary fault. The maximum subsidence was approximately 350 m with a resulting collapse volume of  $2 \text{ km}^3$  (Filson et al., 1973; Francis, 1974).

Minor seismic activity ( $M_s < 4.2$ ) was recorded during the first week of June (Fig. 5.27B). On June 12, a sharp increase in seismic activity began and reached its maximum on June 19. Then, the number of earthquakes declined rapidly, and the swarm practically stopped after June 23. This seismic swarm was characterized by a great number of high-magnitude events that had

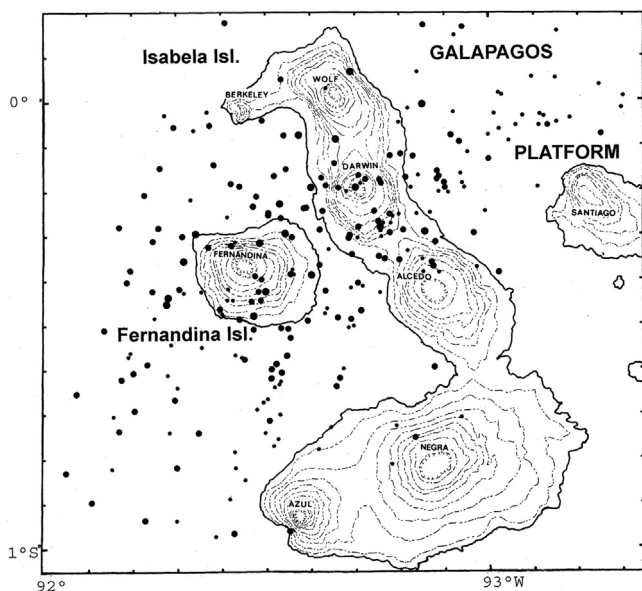


**FIGURE 5.26** Epicenters of the 1968 Galapagos swarm earthquakes located by NOS network. Large distance of the nearest stations from the epicenters (about 400 km) produced a large error in locations; therefore, the great area of distribution of earthquakes may not reflect the real situation. From Filson, J., Simkin, T., Leu, L.-K., 1973. Seismicity of a caldera collapse: Galapagos Islands 1968. *Journal of Geophysical Research* 78, 8591–8622.

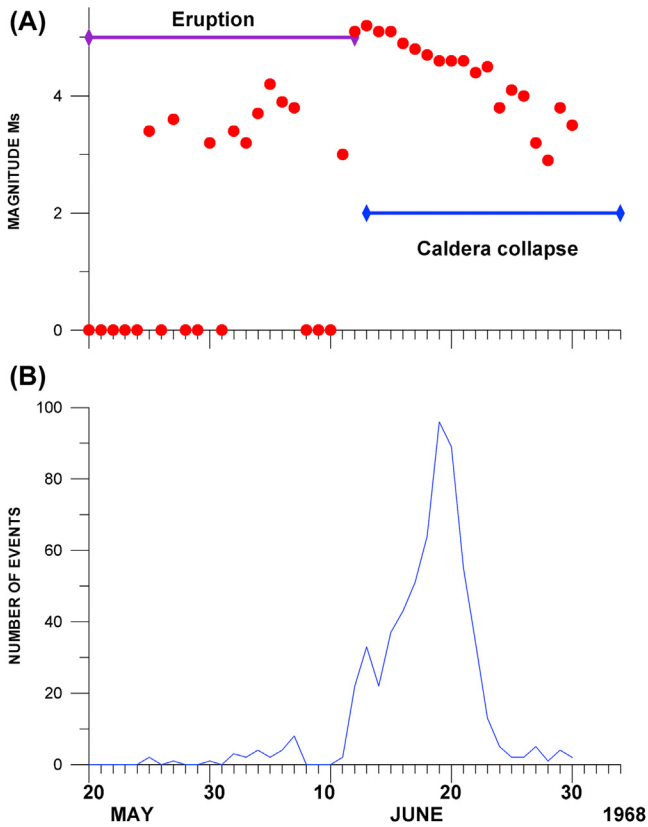


**FIGURE 5.27** Distribution of maximum magnitudes,  $M_s$  (A) and daily number of earthquakes (B) from May 20, 1968 to June 30, 1968. From Filson, J., Simkin, T., Leu, L.-K., 1973. *Seismicity of a caldera collapse: Galapagos Islands 1968. Journal of Geophysical Research* 78, 8591–8622.

practically the same form (Filson et al., 1973; Kaufman and Burdick, 1980). Seismic activity developed in two stages. The first stage was associated with the eruptive activity of the volcano (May 20–June 11). The maximum magnitude of events recorded during this stage was 4.2. The second stage began on June 12, when the magnitude and number of events began to sharply increase. The maximum daily magnitudes 5.1–5.2 were recorded during 4 days, that is, June 12 to June 15 (Fig. 5.27A). The total number of events with magnitude  $M_s \geq 5.0$  was 5 (Filson et al., 1973), which is very large for any volcanic earthquake swarm. Filson et al. (1973) suggested that the caldera collapse occurred during the second stage of seismic activity, and the earthquakes resulted from the dropping of the caldera block. All epicenter locations were made by the global seismic network (Fig. 5.26), which

produced large errors in location. It made it difficult to do any tracing of epicenters and comparison with the volcanic structures.

## 5.5 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH SUBMARINE ERUPTIONS

Submarine eruptions vary between its birth on the deep sea floor and its emergence above sea level. Evidence of deep submarine eruptions is rarely visible at the surface because the steam they produce is quickly condensed as it rises through cold seawater. At the same time, if the eruption is vigorous and the vent is within 500 m or so to the surface, the water may become turbulent and cloudy (Bardintzeff and McBirney, 1999). The best-documented seismological examples are two eruptions that occurred on the volcanic chain of Izu islands, Japan: the 1989 Teishi Knoll eruption and the 2000 Miyakejima eruption and the 2010–11 submarine eruption of El Hierro volcano in the Canary Islands.

### 5.5.1 Teishi Knoll Volcano, Izu Islands

On June 30, 1989, an earthquake swarm started off the eastern Izu Peninsula in Japan (Figs. 5.28 and 5.29). This was then culminated in the submarine

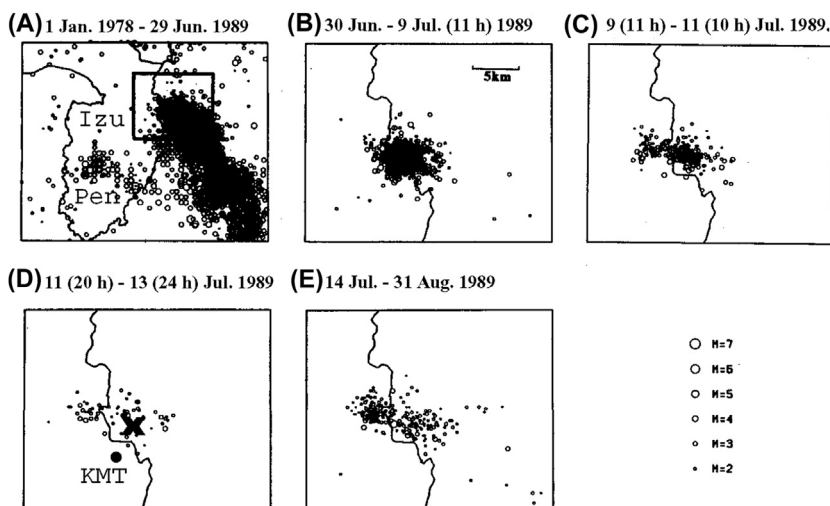


**FIGURE 5.28** Spatial-temporal development of the 1989 Teishi Knoll earthquake swarm. (A) Seismic activity of the zone of eruption before the swarm, (B) seismicity during stages I and II of the swarm, (C) seismicity during stage III of the swarm, (D) seismicity just before the eruption, and (E) seismicity during the eruption. Inset in (A) shows the area of study in (B–E). Cross in (D) shows the site of eruption. KMT is a seismic station. From Yamasato, H., Yokota, T., Kashiwabara, S., 1991. Earthquake swarm and volcanic tremors off Eastern Izu Peninsula in 1989. Spectral investigation and characteristics of waveforms. *Journal of Physics of the Earth* 39, 79–92.



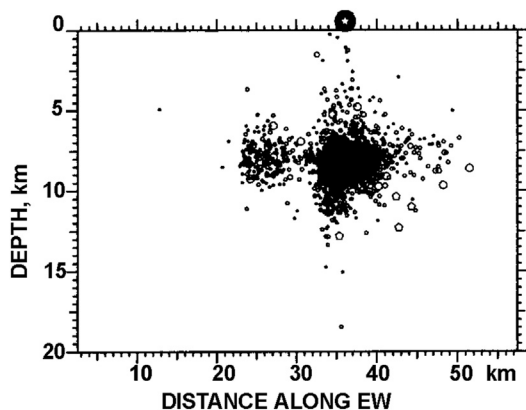


**FIGURE 5.29** Depth distribution of the 1989 Teishi Knoll earthquake swarm events along the EW direction. Black circle shows the site of the eruption. From Matsumura, S., Ohkubo, T., Imoto, M., 1991. Seismic swarm activity in and around the Izu Peninsula preceding the volcanic eruption of July 13, 1989. *Journal of Physics of the Earth* 39, 93–106.

eruption on July 13. Hourly numbers of earthquakes (Fig. 5.30) that were observed at the station KMT, situated at a distance of 6.8 km from the eruption site, show (Fig. 5.28; Yamasato et al., 1991) that three main stages in the development of the swarm were observed: (1) from June 30 to July 4; (2) from July 4 to July 8; and (3) from July 8 to July 13.

During stage I, seismic activity was very low; a great increase in the number of events began on July 4, and then gradually decreased during the next 3 days (stage II). The next large impulse in seismic activity (stage III) began on July 8 and reached its maximum on July 9, when the  $M$  5.5 earthquake, the largest in this sequence, occurred. Then, the seismic activity gradually declined and the eruption began on July 13 with a low background level.

Fig. 5.28 demonstrates that regularization of the epicentral zone of events during the swarm was observed. During stages I and II, the epicentral zone had an elliptical form (Fig. 5.28B). At stage III, the epicenters clustered in a lineal submeridional zone including the site of forthcoming eruption (Fig. 5.28C and D). After the beginning of the eruption, seismic events continued to cluster within this lineal zone. The majority of seismic events of the swarm were located within the depth range from 5 to 10 km. Two hypocentral subzones were formed: the initial large subzone just below the eruption site that was active during stages I and II, and a narrow smaller size subzone that was formed during stage III of the swarm activity. A thin lineal zone of a few events connects the large hypocentral subzone with the site of eruption (Fig. 5.29).

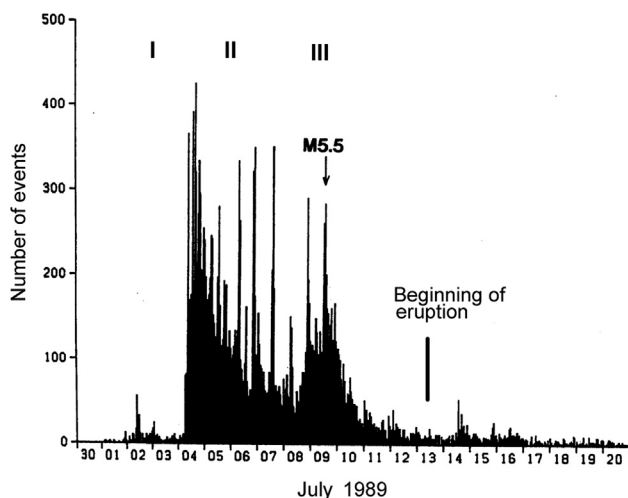


**FIGURE 5.30** Temporal variations of the number of earthquakes during the 1989 Teishi Knoll swarm according to the data of the station KMT (see Fig. 5.28D). From Yamasato, H., Yokota, T., Kashiwabara, S., 1991. *Earthquake swarm and volcanic tremors off Eastern Izu Peninsula in 1989. Spectral investigation and characteristics of waveforms. Journal of Physics of the Earth* 39, 79–92.

## 5.5.2 Miyakejima Volcano, Izu Islands

Miyakejima is an active volcanic island consisting of a basaltic stratovolcano with a 9-km basal diameter and a maximum elevation of about 750 m (Fig. 5.31). The volcanic island has been active for more than 10,000 years. After 17 years of quiescence, on the evening of June 26, 2000, small volcanic earthquakes began to be recorded at the stations west of the summit. Seismic activity developed into two successive swarms of volcano-tectonic earthquakes associated with two eruptive events: the flank submarine eruption and the following caldera collapse in the crater of the volcano. The swarms continued for about 3 months. Fig. 5.31 shows the main seismovolcanic elements of the eruption: the site of the submarine eruption, the area of the collapsed caldera, and the area of the earthquake swarm. The development of the June 2000 to August 2000 seismic activity is shown in Figs. 5.31 and 5.32 (Geshi et al., 2002; Nakada et al., 2005).

### 5.5.2.1 Seismicity Associated With the Submarine Eruption

The first swarm of volcano-tectonic earthquakes began beneath the summit at depths from 1 to 4 km (Fig. 5.32A) on June 26, 2000, but the summit eruption did not occur. Hypocenters of earthquakes moved westward indicating an intrusion of a dyke oriented ENE–WSW and originating at the west coast of Miyakejima. A flank submarine eruption (Fig. 5.33) occurred on the morning of June 27, coinciding with a sharp increase in the number of earthquakes

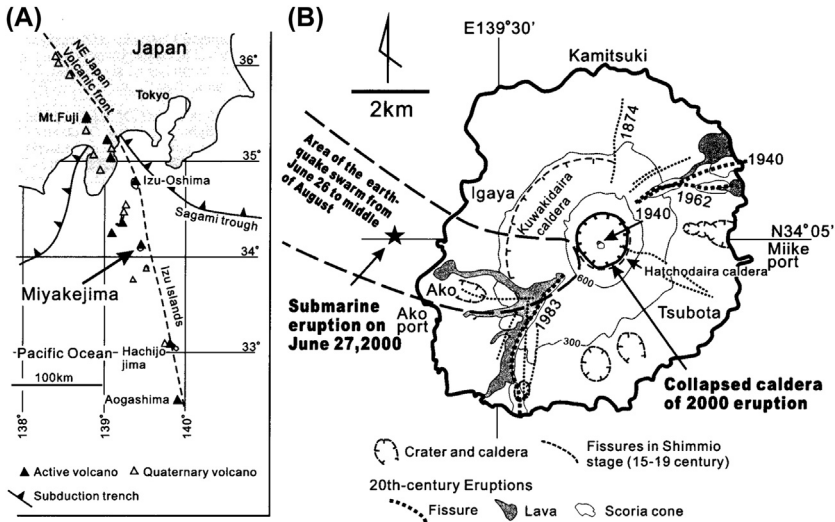


**FIGURE 5.31** The position of Miyakejima volcano in Izu islands (A) and the main seismovolcanic elements of the 2000 eruption (B). Outline of the caldera formed during the 2000 eruption, the location of submarine eruption on June 27, and the approximate area of the earthquake swarm during July and August are shown. From Geshi, N., Shimano, T., Chiba, T., Nakada, S., 2002. Caldera collapse during the 2000 eruption of Miyakejima volcano, Japan. *Bulletin of Volcanology* 64, 55–68.

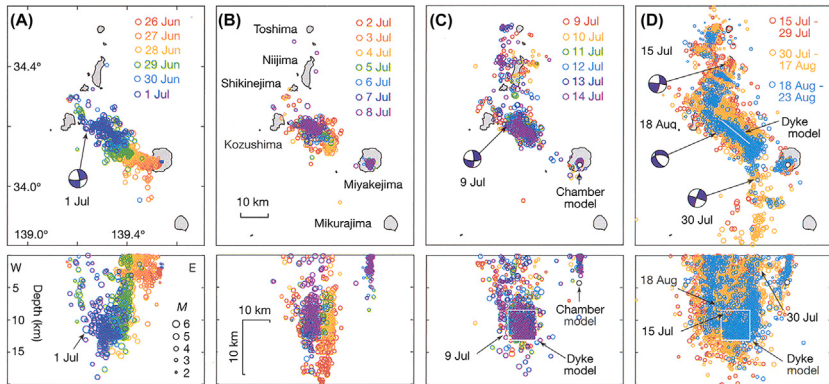


**FIGURE 5.32** The 2000 Miyakejima earthquake swarm evolution in map view (top panels) and cross-sections (bottom panels) during the periods from June 26 to July 1 (A), from July 2 to 8 (B), from July 9 to 14 (C) and from July 15 to August 23 (D). Arrows show the epicenters and hypocenters of the  $M_w > 5.7$  earthquakes (July 1,  $M_w$  6.2; July 9,  $M_w$  5.9; July 15,  $M_w$  6.0; July 30,  $M_w$  6.4; and August 18,  $M_w$  5.7). Their focal mechanism solutions are shown also. Earthquake magnitudes were estimated by NIED, Tsukuba. From Toda, S., Stein, R.S., Sagiya, T., 2002. Evidence from the AD 2000 Izu islands earthquake swarm that stressing rate governs seismicity. *Nature* 419, 58–61.

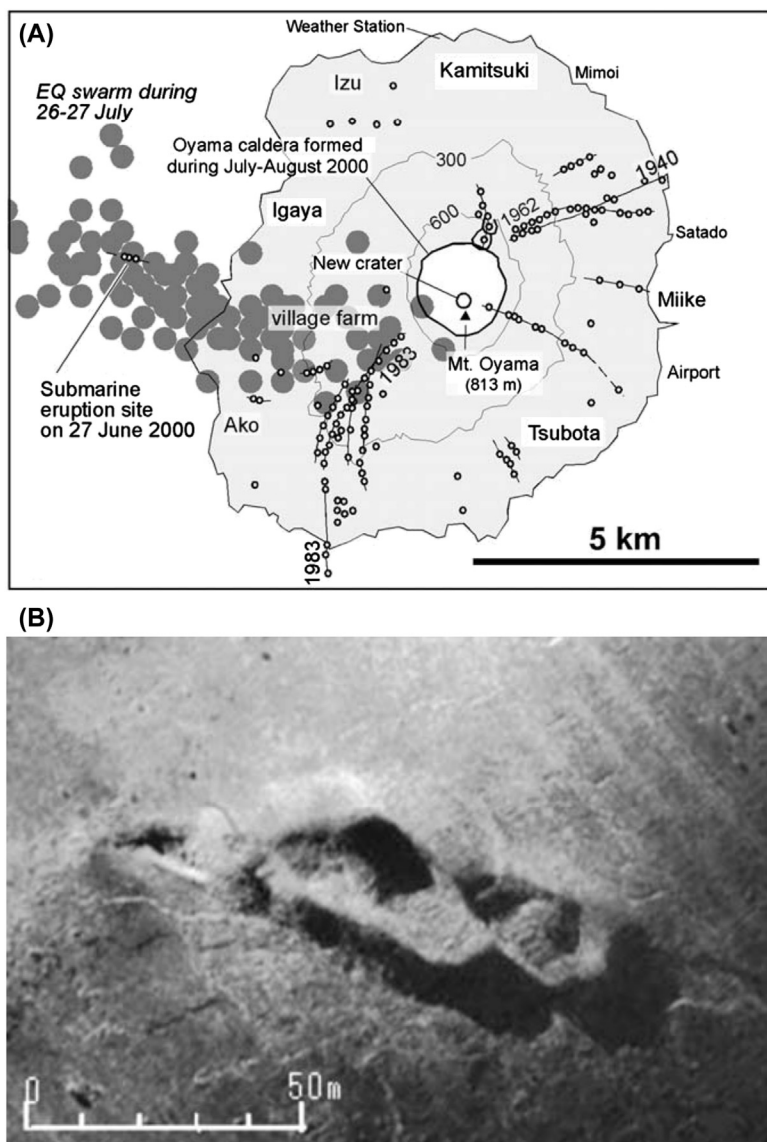


**FIGURE 5.33** Epicenters of the Miyakejima earthquakes. (filled circles) occurring from June 26, 2000, to June 27, 2000, and the submarine eruption site (A) Craters of old eruptions and new submarine craters are shown by small open circles. Radial dyke patterns are shown also by lines with small open circles. The multibeam sides scan sonar images of the new craters formed during this eruption are shown in (B). From Nakada, S., Nagai, M., Kaneko, T., Nozawa, A., Suzuki-Kamata, K., 2005. Chronology and products of the 2000 eruption of Miyakejima volcano, Japan. *Bulletin of Volcanology* 67, 205–218; Kaneko, T., Yasuda, A., Shimano, T., Nakada, S., Fujii, T., Kanazawa, T., Nishizawa, A., Matsumoto, Y., 2005. Submarine flank eruption preceding caldera subsidence during the 2000 eruption of Miyakejima volcano, Japan. *Bulletin of Volcanology* 67, 243–253.

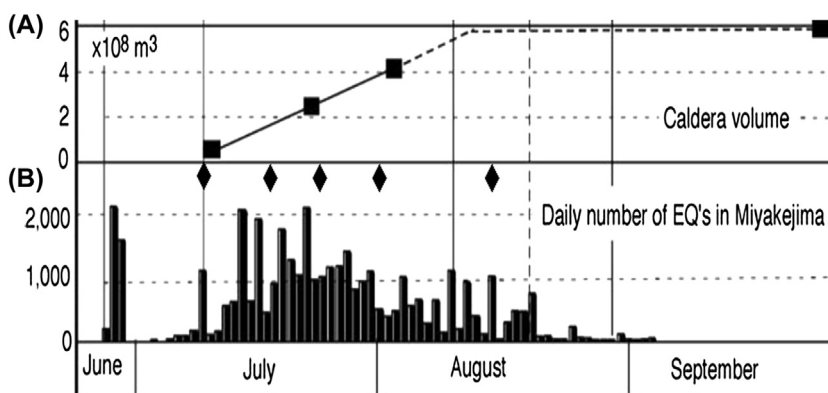


**FIGURE 5.34** The rate of caldera volume growth (A) and the daily variations in the number of the Miyakejima earthquakes (B). Appearance of large ( $M > 6$ ) earthquakes is shown by black diamonds. From Nakada, S., Nagai, M., Kaneko, T., Nozawa, A., Suzuki-Kamata, K., 2005. Chronology and products of the 2000 eruption of Miyakejima volcano, Japan. *Bulletin of Volcanology* 67, 205–218.

(Fig. 5.34), probably soon after the seismic swarm passed beneath the eruption site. Four craters of 20–30 m diameter aligned EW in about  $1500 \text{ m}^2$  area on the sea floor were revealed (Fig. 5.32B; Kaneko et al., 2005; Nakada et al., 2005).

The earthquake activity continued westward filling the space between Miyakejima and Kozushima islands and between the depths from 0 to 15 km (Fig. 5.32A). The activity resumed on June 28 (Fig. 5.33). The single  $M_w$  6.2 (NIED) earthquake occurred on the western end of the epicentral area on July 1.

### 5.5.2.2 Seismicity Associated With the Caldera Collapse

On July 8, the collapse of the summit area suddenly began being accompanied by a small explosion. Subsidence of the summit area continued through early August, simultaneously with the intense second earthquake swarm. The swarm occurred at depths between 0 and 20 km located between Miyakejima and Kozushima islands and to the north of Kozushima island (Fig. 5.32B–D) and included four strike-slip  $M_w > 5.7$  (NIED) earthquakes. The July 9 earthquake ( $M_w$  5.9) occurred at the same place as the July 1 event from the first swarm. Three other earthquakes occurred at the ends of the swarm area ( $M_w$  6.0, July 15 and  $M_w$  6.4, July 30) and in its central zone ( $M_w$  5.7, August 30). The swarm gradually declined in the second half of August (Nakada et al., 2005; Uhira et al., 2005). A collapsed caldera, 1.6 km in diameter and 450 m in depth, was formed. Its final volume was about  $6 \times 10^8 \text{ m}^3$  (Geshi et al., 2002). During subsidence, phreatomagmatic explosions occurred on July 14 and July 15. Nakada et al. (2005) suppose that the subsidence occurred due to lateral migration of magma under the ground.

Toda et al. (2002) discussed the model of seismicity compatible with a laterally propagating dyke intrusion during this eruption. They assumed that a

vertical dyke propagated during the first week (Fig. 5.32A) and then opened continuously for 7 weeks. The geometry of the inferred dyke proposed by Toda et al. (2002) is shown in Fig. 5.32C and D. They consider that the process of dyke expansion and associated “dog-bone” pattern of seismicity was controlled by stress transferred from an expanding dyke to vertical faults in the surrounding crust with the mean stressing rate of 32 bar/year.

### 5.5.3 El Hierro Volcano, Canary Islands

El Hierro island is the emergent summit of a volcanic shield that rises from 3800 to 4000 m depth and grows up to 1500 m a.s.l. It is formed by different volcanic edifices and includes three rift zones along which volcanism is concentrated. El Hierro had no confirmed eruptions during the last 500 years; some uncertainty exists on a possible event in 1793 (Ibáñez et al., 2012).

The 2011 submarine eruption took place about 2 km off the southern coast of El Hierro (). The eruption of basanitic magma constructed a submarine cone at a depth of about 300 m. This cone reached a height of 220 m with a basal diameter of over 1 km. The total amount of erupted material was of the order of 0.33 km<sup>3</sup> (Rivera et al., 2013). Unrest was first detected in July 2011, when the seismic network of the Instituto Geográfico Nacional (IGN) began locating a large number of relatively small volcano-tectonic earthquakes ( $ML < 2.5$ ) underneath El Hierro island, at depths between 8 and 15 km. The eruption started on October 10, 2011, and continued till March 5, 2012.

During the period from July 2011 to March 2012, the seismicity developed within three swarms well separated in time and space as is seen in Figs. 5.35 and 5.36. These figures are based on the catalog of earthquakes prepared by the IGN. Fig. 5.35, representing the distribution of epicenters and hypocenters of the earthquakes, is modified from Ibáñez et al. (2012). Ibáñez et al. (2012) showed that the magnitude of completeness of the IGN catalog varied for different stages of the seismic activity between  $ML$  1.3 and 1.9. To have a representative view of the temporal variations of the daily numbers and maximum magnitudes of the events, Fig. 5.36 demonstrates these characteristics are taken from the IGN catalog for events with  $ML \geq 2.0$ . The following description of three volcano-tectonic earthquake swarms and their nature is based on the data shown in Figs. 5.35 and 5.36, and the results of the analysis of associated surface deformation and petrological data are presented by Martí et al. (2013).

The first, preliminary, earthquake swarm was recorded from July 20, 2011, to September 20, 2011. The events of this swarm were located within the southern part of the El Golfo depression and underneath El Hierro island, spreading across the western rift zone at depths of about 8–13 km. They were low magnitude (only for two events magnitude reached 2.7); the daily number of events with  $ML \geq 2.0$  varied between 0 and 27. Martí et al. (2013) consider that this swarm reflected the uplift of a batch of basanitic magma coming from a reservoir located at a depth of about 25 km with the creation of a

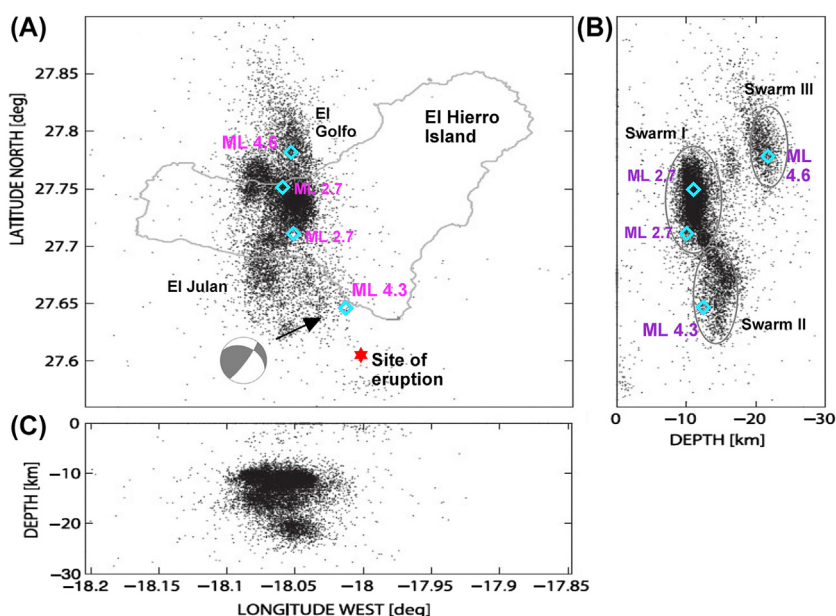


**FIGURE 5.35** Epicenters (A) and hypocenters (B and C) of three swarms (I, II, III) of volcano-tectonic earthquakes associated with the 2011–12 submarine eruption of the El Hierro volcano (Canary Islands). The epicenters and hypocenters of the events with maximum magnitudes of each swarm are shown. The focal mechanism solution for the  $ML$  4.3 event of October 8, 2011, is shown according to [Del Fresno et al. \(2015\)](#). Ellipses limit the hypocentral areas of the three swarms. Modified from [Ibáñez, J.M., De Angelis, S., Díaz-Moreno, H.P.A., Alguacil, G., Posadas, A., Pérez, N., 2012. Insights into the 2011–2012 submarine eruption off the coast of El Hierro \(Canary Islands, Spain\) from statistical analyses of earthquake activity. \*Geophysical Journal International\* 191, 659–670.](#)

shallower magma reservoir and a magma-driven fracturing of the crust necessary to create a space to accumulate magma at this shallower position.

The second, preceding, swarm occurred between September 20, 2011, and October 22, 2011. It clustered to the south from the swarm I events, mainly beneath the sea floor of El Julian Bay at depths ranging between 10 and 20 km and reflecting, as considered by [Martí et al. \(2013\)](#), the lateral migration of magma toward the SE. A swarm began with a sharp increase in the number of events reaching 105–140 daily earthquakes from September 27 to September 29. Then, the frequency of the earthquakes began to gradually decrease to 39 events with  $ML \geq 2.0$  recorded on October 8. At the same time, the maximum daily magnitude of earthquakes increased, and the largest preeruption earthquake with magnitude  $ML = 4.3$  ( $M_w = 4.0$ ) occurred on October 8 at a depth of 12 km. The source mechanism of this event indicated a strike-slip motion with a reverse component and fault planes oriented in NE–SW and WNW–ESE directions ([Del Fresno et al., 2015](#)). During the next 2 days, up to the beginning of the eruption on



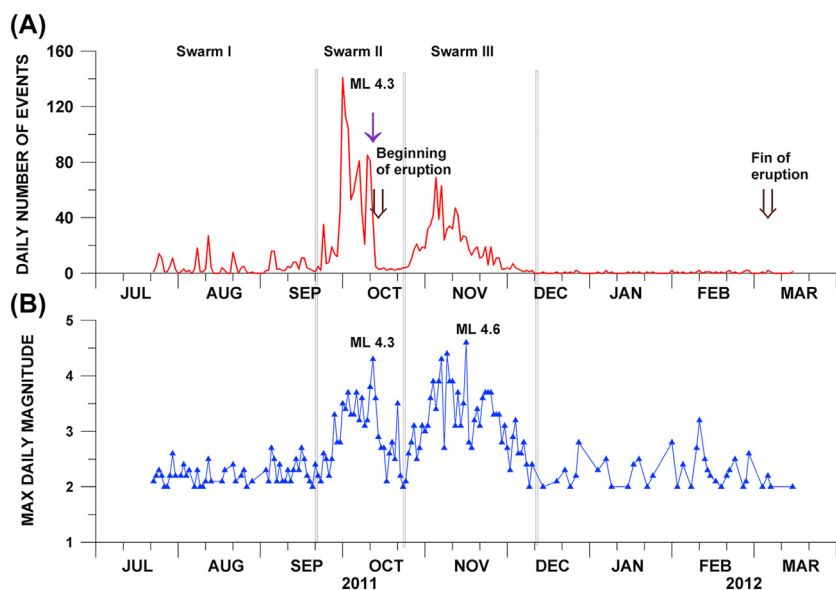


**FIGURE 5.36** Temporal variations in the daily number of volcano-tectonic earthquakes with magnitude,  $ML \geq 2$  associated with the submarine eruption of the El Hierro volcano (A) and the maximum daily magnitudes of these events (B). Vertical lines separated the appearance of the three swarms of earthquakes. A thin arrow shows the moment of appearance of the largest earthquake from the preceding swarm. Double arrows show the beginning and the end of the eruption. Data from the catalog of Instituto Geográfico Nacional.

October 10, the daily number of events varied between 3 and 5. The eruption began on the background of low-level volcano-tectonic seismicity and volcanic tremor, 5 km away from the epicenter of the  $ML$  4.3 earthquake.

The third, continuous, swarm of volcano-tectonic earthquakes occurred during the eruption period but out of the eruption area, to the north from the swarm I. The earthquakes of this swarm were located beneath the El Golfo depression (Fig. 5.35). The swarm began on October 22, 12 days after the eruption onset, and continued till the first days of December 2011 simultaneously with the low-level seismicity in the area of eruption. It was a rather deep seismic sequence recorded at depths between 18 and 25 km. The swarm was characterized by a gradual increase in the daily number and magnitude of events till November 11 when the largest earthquake with  $ML = 4.6$  occurred. Then, a gradual decrease of these parameters was observed (Fig. 5.36).

The swarm III, as Ibáñez et al. (2012) wrote, stands out almost as an independent earthquake sequence. The appearance of this swarm beneath the El Golfo depression, 20–30 km to the north of the area of the eruption, with increase of focal depths of earthquakes and their magnitudes up to  $ML$  4.6, may be result of the subsidence of the crustal structures covering the initial



magma chamber, that is caused, as supposed Martí et al. (2013), by readjustments of the whole plumbing system following the decompression caused by the withdrawal of magma during the eruption. Practically, it was a posteruption earthquake swarm similar to the same observed after the 1975–76 Tolbachik fissure eruption when the seismic activity was recorded during 1977 and 1978 at the periphery of the active volcanic zone at a distance of 40–60 km from the new volcanoes with the magnitudes of the largest posteruption earthquakes reaching 5.2 (see Section 5.3.1).

After the third swarm, low seismic activity, represented by one to two daily seismic events of the magnitude range between 2 and 2.8, continued up to the end of the eruption on March 5, 2012.

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## RELEVANT WEBSITES

- BGVN. Bulletin of the Global Volcanism Network. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.
- GVP. Global Volcanic Program. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.
- Harvard CMT. <http://www.globalcmt.org/CMTsearch.html>.
- USGS. U.S. Geological Survey. <http://earthquake.usgs.gov/>.

## Chapter 6

# Volcano-tectonic Earthquakes at Andesitic Volcanoes: Case Studies

Andesitic volcanoes produce lava with viscosity in the range of  $10^3$ – $10^5$  P at a temperature of about  $1000^\circ\text{C}$ – $1300^\circ\text{C}$ , and the volcanic activity is of explosive and effusive types that are characterized by higher levels of volatiles. Andesitic volcanoes usually have the form of stratovolcanoes. The most famous include Bezymianny in Kamchatka, Merapi in Indonesia, and Volcán de Colima in Mexico.

The eruptions of andesitic volcanoes occur mainly from the summit crater of the volcano and rather seldom on its flanks. The magmas are rich in volatiles, mobile, able to produce phreatomagmatic or phreatic explosions, and can form lava domes in the crater of the volcano. These eruptions are accompanied by lava and pyroclastic flows and numerous rockfalls. The normal cycle for andesitic volcanoes consists of the formation of a lava dome and then its destruction with its collapse or an explosion. The rock fracturing during the formation of the passageway for the magma to the surface produces volcano-tectonic earthquakes. Lava effusion from the summit and flank (fissure) vents may be accompanied by volcanic tremors.

The author discusses the volcano-tectonic seismicity associated with three types of activity of andesitic volcanoes: seismic activity related to the (1) great “directed blasts” at volcanoes, (2) phreatic or phreatomagmatic explosions, and (3) block-lava extrusions. These three types of eruptions may form also subsequent stages of the eruptive activity of an andesitic volcano.

## 6.1 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH VOLCANIC “DIRECTED BLASTS”

The “directed blasts” of andesitic and dacitic volcanoes are within the most “large-scale” eruptive events. During an eruption of this type, a great part of the volcanic cone can be exploded horizontally and end up distributed over great area (Gorshkov and Bogoyavlenskaya, 1965). Typical eruptions are Bezymianny, Kamchatka, 1956; Sheveluch, Kamchatka, 1964; and Mount St. Helens, Cascades, United States, 1980.

### 6.1.1 Bezymianny Volcano, Kamchatka

The March 30, 1956 explosion of the Bezymianny volcano became a classic example of a “directed blast” eruption (Gorshkov and Bogoyavlenskaya, 1965). This stratovolcano with a height of 3085 m was considered as dormant, but on September 29, 1955, the first earthquake in the region of the volcano was recorded. From October 6, a swarm of volcano-tectonic earthquakes (Fig. 6.1) began, and the first explosion was recorded on October 22, 1955. Strong ash eruption continued for 1 month, but it changed to moderate explosive activity that culminated in the paroxysmal explosion of March 30, 1956. This explosion destroyed the SE part of the volcanic cone and pyroclastic material was erupted to the SE at the angle of 30–40 degrees to the horizon. Ash cloud rose up to 40 km.

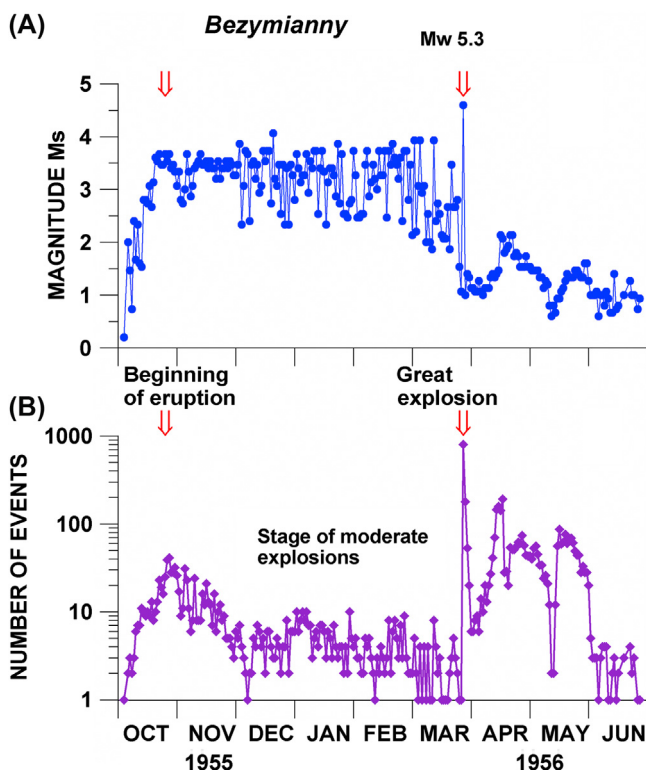


**FIGURE 6.1** Daily distribution of maximum magnitudes (A) and the number of earthquakes (B) associated with the October 1955 to June 1956 eruption of Bezymianny volcano (station KLY, 42 km from Bezymianny volcano). Arrows show the beginning of the eruption, the appearance of the great directed explosion of March 30, 1956 and the moment of magnitude  $M_w$  5.3 earthquake. Data from Tokarev, P.I., 1981. *Volcanic Earthquakes of Kamchatka*. Nauka, Moscow (in Russian). 164 p.

An area of  $500 \text{ km}^2$  at a distance of 30–32 km to the SE of the volcano was covered by about  $3 \text{ km}^3$  of erupted material. The height of the volcano was diminished by about 180 m. Later, in April 1956, a new dome began to grow in the crater. This process of the formation and destruction of the dome continued for many years, after 1965 accompanied by extrusions of block lava (Bogoyavlenskaya and Kirsanov, 1981; Bogoyavlenskaya et al., 1991).

Fig. 6.1 shows the development of seismic activity during 1955–56. The only seismic station KLY was situated at a distance of 42 km from the volcano and was equipped with short and mediate period three-component seismometers with a magnification of  $\times 5000$  and  $\times 300$ , respectively. Although it failed to localize the events, it was used to count the number of events and to estimate their magnitude. One month before the eruption, the number and energy release of seismic events gradually increased up to the first explosion. During the next stage, when moderate explosions were observed, the number of events gradually decreased during November and stayed at a stable low level (about 10 events daily) from December 1955 to March 1956. The daily earthquake magnitude level during this stage was rather high, reaching 3.5. The paroxysmal explosion of March 30 occurred in the background of the low level of seismic activity and was associated with the magnitude  $M_w$  5.3 earthquake. The postexplosion stage of dome building was characterized by a gradual decrease in the number and magnitude of seismic events.

Gorshkov and Bogoyavlenskaya (1965) noted that the variations in the number of seismic events correlated well with the manifestations of eruptive activity while the variation in the maximum magnitude of earthquakes had no such dependence. Larger earthquakes formed a permanent background from October 18, 1955 (before the eruption) to the end of February 1956, but their occurrence did not depend on the eruptive manifestations.

After the paroxysmal explosion, the growth of a lava dome in the crater was observed. The partial destruction of the dome by explosions was preceded by an increase in seismic activity 2–5 days before the eruption (Tokarev, 1966, 1981; Bogoyavlenskaya et al., 1991).

Two typical variations of seismic activity before the explosions of the dome during 1959 and 1960 are shown in Fig. 6.2. The number of events was counted at seismic station KLY (Tokarev, 1981). During the 1959 event, an increase in seismic activity began on March 26 (Fig. 6.2A). The number of events sharply increased up to the moment of the March 28 explosion. After the explosion, the number of events gradually decreased. The majority of seismic events had magnitudes from 0 to 1. The maximum event with a magnitude of 1.3 was recorded on March 28.

A similar situation was observed before the April 13, 1960 to April 14, 1960 eruption (Fig. 6.2B). An increase in seismic activity began 3 days before the explosion; the explosion occurred at the maximum level of

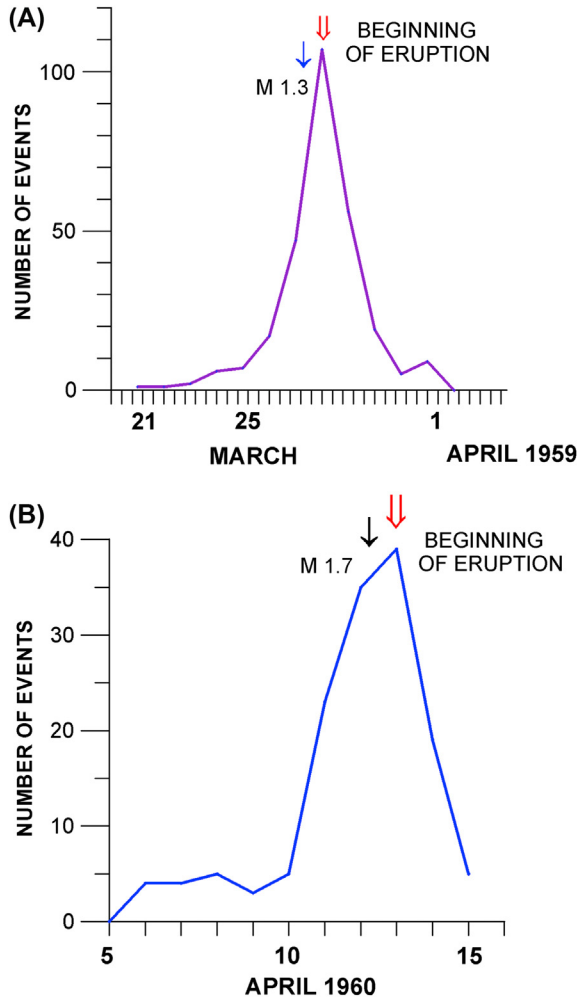


**FIGURE 6.2** Variations in the number of earthquakes before two explosions of Bezmianny volcano that occurred during 1959 (A) and 1960 (B) and were observed at KLY station at a distance of 42 km from the Bezmianny volcano. Data from Tokarev, P.I., 1981. *Volcanic Earthquakes of Kamchatka. Nauka, Moscow (in Russian)*. 164 p.

seismic activity. After the explosion, the number of events gradually decreased. The majority of seismic events had magnitudes from 0 to 1. The maximum event with a magnitude of 1.7 was recorded on April 12.

### 6.1.2 Sheveluch Volcano, Kamchatka

Sheveluch volcano is the northernmost active andesitic volcano of Kamchatka. It is located at the junction of the Kurile-Kamchatka and

Aleutian subduction zones. It consists of two parts: the old edifice with an altitude of 3283 m and the young (active) edifice with an altitude of 2800 m. During the last 10,000 years, the volcano has been characterized by two types of eruptions: the great eruptions of the “directed blast” type (they have continued from hours to a few days) and the slight to moderate eruptions associated with the growth and destruction of the crater dome (they have continued for several years) (Bogoyavlenskaya et al., 1991).

### 6.1.2.1 1964 Eruption

The “directed blast” eruption took place on November 12, 1964, after 15 years of calm. It continued for only 70 minutes. As a result, about 1.5 km<sup>3</sup> of resurgent material and 0.8 km<sup>3</sup> of juvenile material were ejected, and the summit crater and five periphery domes were destroyed by the explosion. A new crater was formed. Fig. 6.3 shows the seismic history of the eruptive process (Tokarev, 1967; Gorshkov and Dubik, 1970).

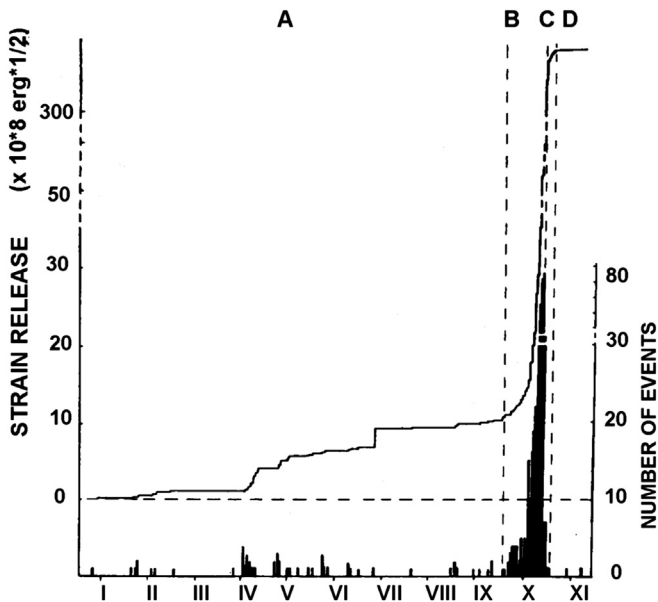


**FIGURE 6.3** Characteristics of seismic activity at Sheveluch volcano in 1964. The strain release during the seismic activity (as a cumulative value of the quadratic root of seismic energy, in Erg) and variations in the number of earthquakes according to the records at seismic station KLY situated at a distance of 45 km to the south of the volcano are shown. The stages in seismovolcanic activity: (A) the preliminary stage (January to the middle of October 1964), (B) the preliminary seismic swarm (October 17 to November 12), (C) the paroxysmal eruption on November 12 and (D) the posteruptive stage. From Tokarev, P.I., 1967. *Gigantic eruption of Sheveluch volcano and its precursors*. *Izv. USSR Ac. Sci., Fizika Zemli* 9, 11–22 (in Russian).



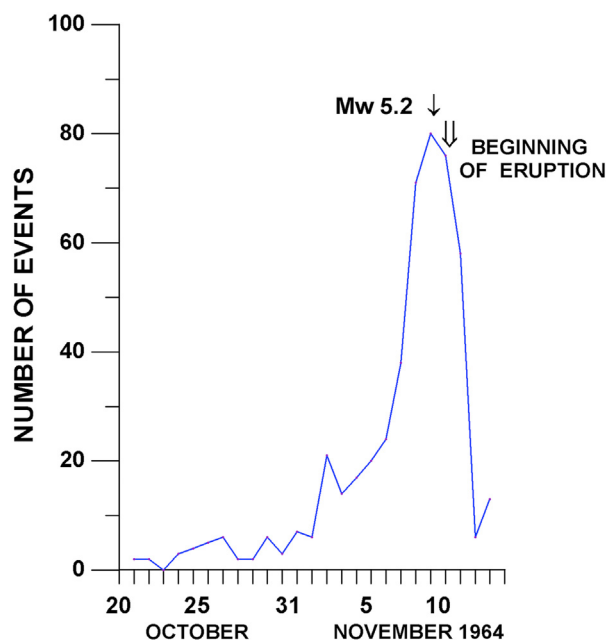


**FIGURE 6.4** Variations in the number of earthquakes before the November 11, 1964 eruption of Sheveluch volcano. Data from the Kamchatka Regional Earthquake Catalog.

The first alert pronouncing the volcanic unrest was initiated by the April 1964–May 1964 swarm of 37 small shallow earthquakes located near the volcano. This swarm was not followed by any volcanic manifestations. The next swarm of volcano-tectonic earthquakes began on October 17 (Fig. 6.4). About 500 earthquakes with magnitude  $M \geq 2$  were recorded. Epicenters of all the earthquakes were located by three seismic stations of Klyuchevskaya Volcano Station and occurred within the volcanic edifice at depths of less than 5 km. The number and energy of earthquakes (according to the data of the nearest seismic station KLY situated at a distance of 45 km to the south of the volcano) gradually increased up to the moment of the paroxysmal explosion of November 12. The maximum magnitude earthquakes were associated with the November 12 paroxysmal eruption; their magnitudes  $M_w$  were 5–5.2 (Tokarev, 1967; Zobin, 2001). The airwaves of the November 12 explosion were recorded by barographs and microbarographs at a distance of up to a few thousand kilometers. After the November 12 explosion, seismic activity sharply decreased and ceased in 4 days.

#### 6.1.2.2 1993 Eruption

A new dome began to grow during the 1980s within the crater that was formed by the 1964 explosion, without any preceding seismic events. Its 1980–93

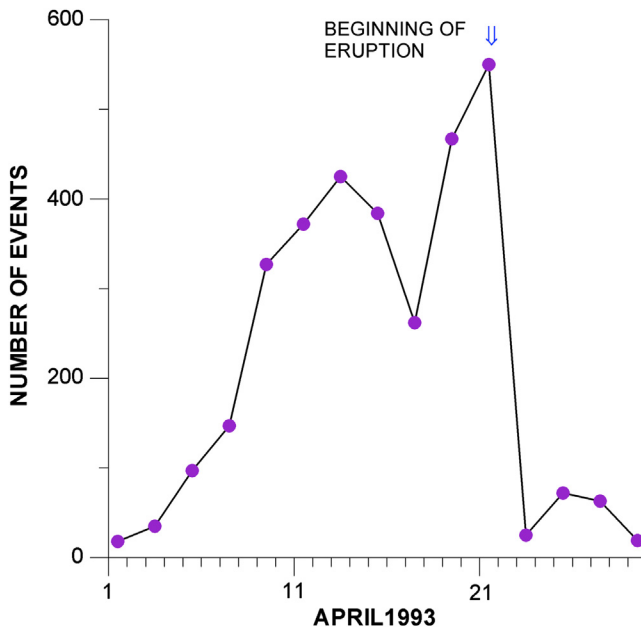


**FIGURE 6.5** Variations in the 2-day number of earthquakes before the 1993 eruption of Sheveluch volcano according to the records of seismic station SVL situated at a distance of 8 km from the crater. A double arrow shows the beginning of the eruption. From Gorelchik, V.I., Shirokov, V.A., Firstov, P.P., Chubarova, O.S., 1997. Shiveluch volcano: seismicity, deep structure and forecasting eruptions (Kamchatka). *Journal of Volcanology and Geothermal Research* 78, 121–132.

growth was accompanied by infrequent Vulcanian and phreatic explosions and culminated in the April 21, 1993 paroxysmal explosion of the dome that produced about  $0.5 \text{ km}^3$  of pyroclastic material (Gorelchik et al., 1995, 1997).

Fig. 6.5 shows the temporal variations in the number of earthquakes preceding the eruption. A seismic swarm began 3 weeks before the eruption. The majority of earthquakes occurred within a distance of less than 5 km from the crater dome; their depths did not exceed 3 km below sea level (b.s.l.). The number of earthquakes gradually increased up to the moment of the eruption. After the eruption, the number of earthquakes sharply decreased.

## 6.2 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH PHREATIC AND PHREATOMAGMATIC EXPLOSIONS

### 6.2.1 El Chichón Volcano, Mexico

El Chichón volcano is an andesitic stratovolcano in southern Mexico. It erupted in March 1982 and April 1982, after about 600 years of quiescence (Simkin and Siebert, 1994), and produced a sequence of six large

phreatomagmatic explosions, occurring from March 29 to April 4, that destroyed the summit dome and formed a 1-km-wide crater.

The eruption was preceded by a 28-day seismic swarm (Fig. 6.6), which began on March 1, 1982. Seismic activity was recorded by the 14-station seismic network of UNAM that was installed in this region during the period 1979–80 (Havskov et al., 1983; Yokoyama et al., 1992; Zobin and Jiménez, 2008). The swarm began with a sharp increase in the number of earthquakes during the first 5 days of activity. The maximum magnitude event ( $M$  4.0) was recorded on the fifth day of the swarm and was followed by a gradual decrease in seismic activity during the next 2 weeks. The activity increased again on March 19 but practically disappeared about 2 days before the

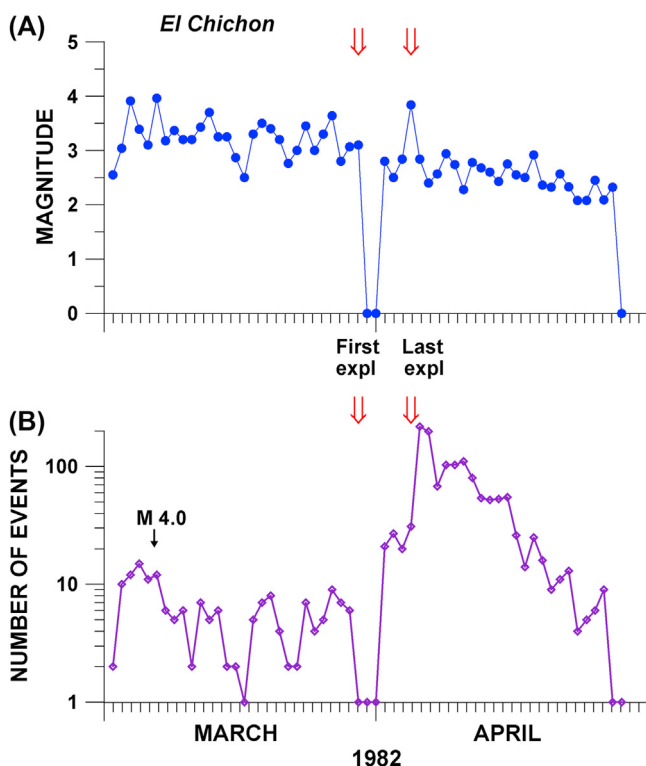


**FIGURE 6.6** Daily distribution of maximum magnitudes (A) and the number of earthquakes (B) associated with the 1982 eruption of El Chichón volcano. Double arrows show the appearance of the first and last explosions, and single arrow shows the moment of the magnitude 4.0 earthquake. From Zobin, V.M., and Jiménez, Z., 2008. Some regularity in the process of re-awakening of andesite and dacite volcanoes: specific features of the 1982 El Chichón volcano, México reactivation. *Journal of Volcanology and Geothermal Research*. 175, 482–487.

eruption. The majority of preceding earthquakes (80%) were located within the area of 10 km  $\times$  20 km with the volcano in the center of this area (Fig. 6.7). They occurred at depths ranging from 2 to 8 km; a small group of foci was located also at depths between 10 and 18 km (Fig. 6.8).

During the explosive activity from March 20 to April 5, the earthquakes clustered densely within the volcano crater zone and to the north of the crater (Fig. 6.7). After the last explosion, seismic activity continued for another 1 month. The depth of events went down up to 20–25 km (Fig. 6.8). The earthquake foci filled the depth interval from 0 to 25 km (Fig. 6.7) and stayed at this depth in April (Jiménez et al., 1999).

## 6.2.2 Volcán De Colima, Mexico

The andesitic stratovolcano Volcán de Colima (height of 3860 m) is the most active volcano in Mexico. The eruptions of Volcán de Colima during the last 480 years involved a wide spectrum of eruption styles, including small phreatic explosions, major block-lava emissions, and large explosive events (Bretón González et al., 2002).

### 6.2.2.1 Phreatic Explosion of 1994

It occurred at 20:13 on July 21. It destroyed the 1991 crater dome and produced avalanches and ash fall. The explosion formed a shallow crater approximately 140 m in diameter and 50 m in depth. The explosion was heard at a distance of 20–35 km to the south of the volcano (Jiménez et al., 1995). The eruption was preceded by a seismic sequence from July 13 to July 21 that was located within the volcanic edifice.

About 450 small seismic events were recorded by the nearest seismic station (the distance from the crater of 1.7 km) during the 1994 swarm. Fig. 6.9 shows the temporal variations in the number of events. The seismic activity developed in two stages: from July 12 to July 15 and from July 17 to July 22. The explosion occurred when the swarm practically finished. The maximum magnitude  $M_w$  1.2 event was recorded during the peak activity of the second stage.

## 6.2.3 Popocatepetl Volcano, Mexico

Popocatepetl volcano is situated within the Trans-Mexican Volcanic Belt. It is a 5452-m-high andesitic stratovolcano. Its most unrest began with a short swarm of volcano-tectonic earthquakes on December 21, 1994, after nearly 70 years of quiescence. A series of moderate phreatic explosions at the crater occurred in about 45 minutes after the beginning of the swarm (De la Cruz-Reyna et al., 1995). The explosions produced emissions of ash which fell on several towns to the east and northeast from the volcano, including the large city of Puebla. This unrest then developed into a sequence of steam-and-ash and fumarolic emissions up to March 1996 when the repetitive dome-

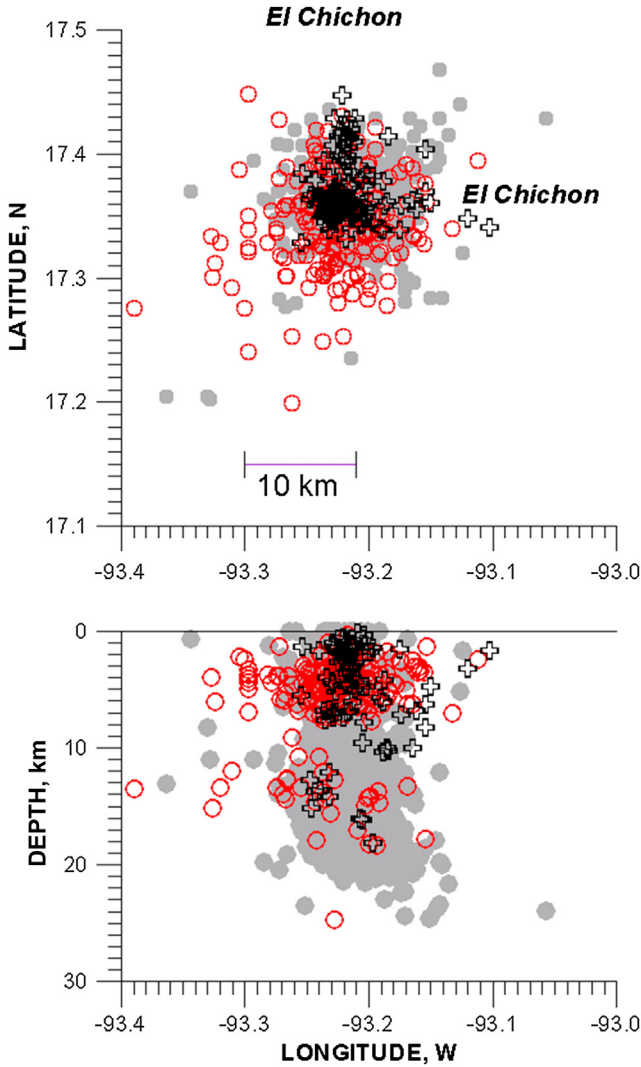


**FIGURE 6.7** Epicenters (above) and hypocenters (below) of the earthquakes associated with the 1982 eruption of El Chichón volcano. *Open circles* indicate the earthquakes that occurred before the first explosion of March 29; *crosses* indicate the earthquakes recorded from March 29 to April 4; *filled circles* indicate the earthquakes that occurred after the explosion of April 4 up to the end of April. The position of El Chichón volcano is shown by the *star*. From Zobin, V.M., and Jiménez, Z., 2008. Some regularity in the process of re-awakening of andesite and dacite volcanoes: specific features of the 1982 El Chichón volcano, México reactivation. *Journal of Volcanology and Geothermal Research*. 175, 482–487.

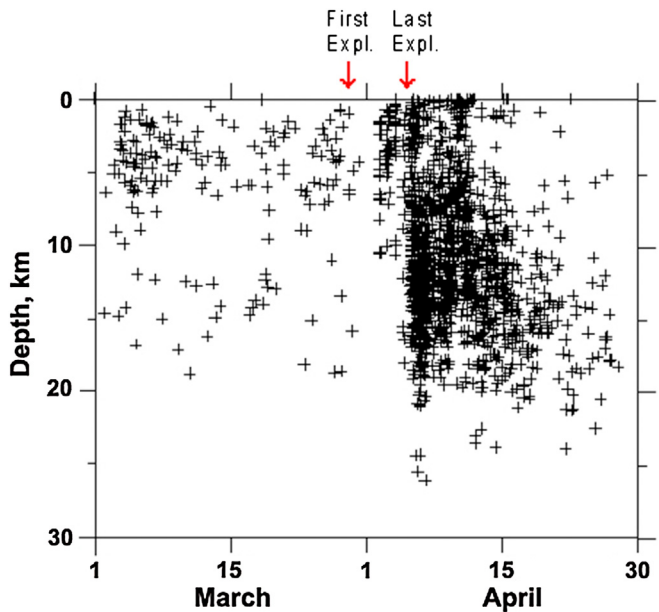


**FIGURE 6.8** Temporal variations of earthquake focal depths during March 1982 and April 1982. The interval between the first and last explosions is shown by arrows. From Jiménez, Z., Espíndola, V.H., Espíndola, J.M., 1999. Evolution of the seismic activity from the 1982 eruption of El Chichón volcano, Chiapas, México. *Bulletin of Volcanology* 61, 411–422.

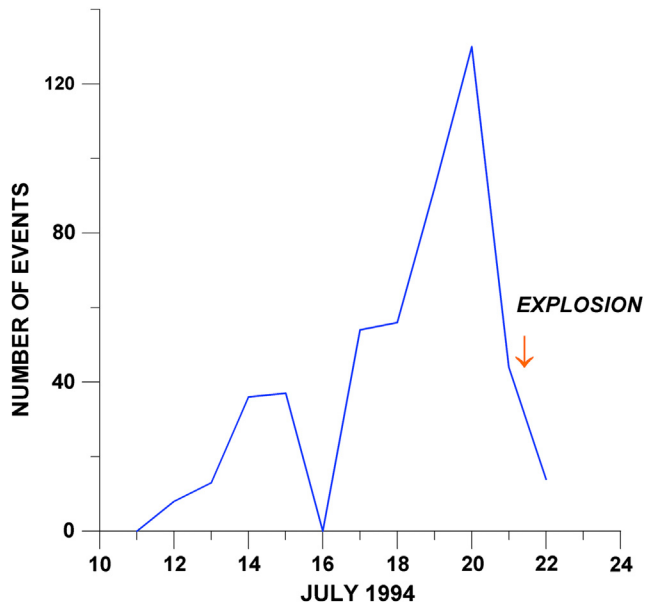


**FIGURE 6.9** Variations in the number of events calculated for every 12-hour intervals before the July 21, 1994 explosion at Volcán de Colima. Data from the Kamchatka Regional Earthquake Catalog.

building and dome-destroying episodes began. Eruptive activity continues up to the present time (De la Cruz-Reyna et al., 2008).

The December 21, 1994 swarm began at 07:03 (GMT; 01:03 LT) on the background of very low seismic activity of the volcano (Fig. 6.10A). Based on the catalog of volcano-tectonic earthquakes, prepared by Lermo-Samaniego et al. (2006), it continued for about 8 hours (Fig. 6.10C) but the majority of earthquakes occurred during the first 45 minutes. The earthquakes were rather small; only 14 volcano-tectonic earthquakes were located by a network

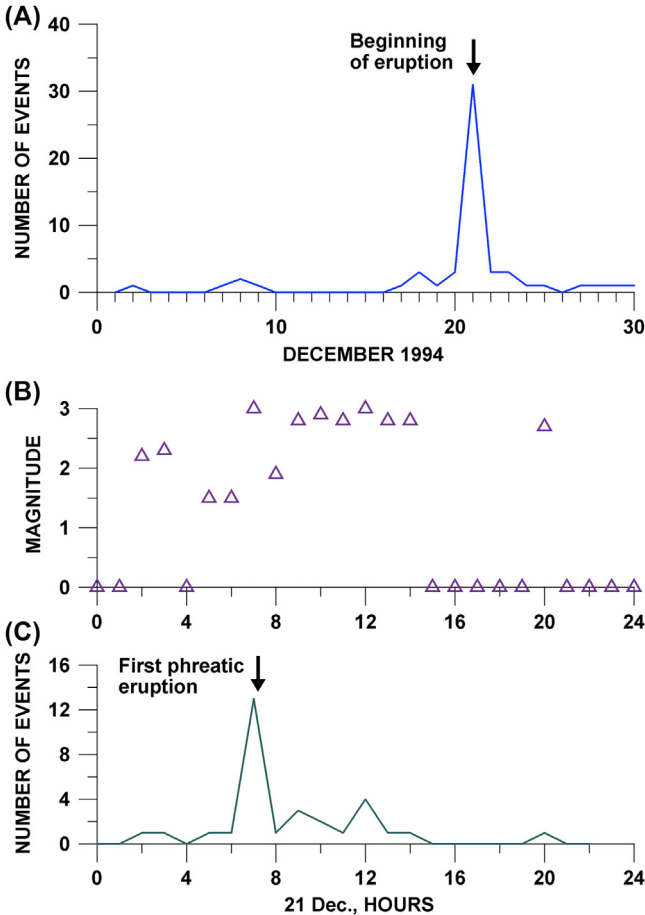


**FIGURE 6.10** Development of the seismic activity at Popocatepetl volcano before and during the December 21, 1994 phreatic eruption. (A) Daily numbers of volcano-tectonic earthquakes recorded during December 1994. Detailed development of the December 21 swarm of volcano-tectonic earthquakes with (B) variations in hourly maximum magnitudes and (C) variations in hourly number of earthquakes. Eruption events are marked with arrows. Data from the catalog of earthquakes prepared by J. Lermo-Samaniego (personal communication, 2003).

consisting of four short-period seismic stations. The maximum magnitude event ( $M$  3.0) occurred about 45 minutes after the beginning of the swarm.

Fig. 6.11 shows the distribution of the seismic events occurring during December 1994. All the earthquakes were located within the volcano edifice. The rare events, recorded before the December 21 crisis, were distributed within the depth range between 2.5 km above sea level (a.s.l.) and 2 km b.s.l. The events of the swarm clustered along the NW–SE direction across the summit area. They occurred at the depth interval between 2 and 5 km a.s.l., reaching the summit elevation. During the last 10 days of December, seismic events occurred at the depths between 0 and 4 km a.s.l. and their epicentral zone spread to SW from the swarm epicenters.

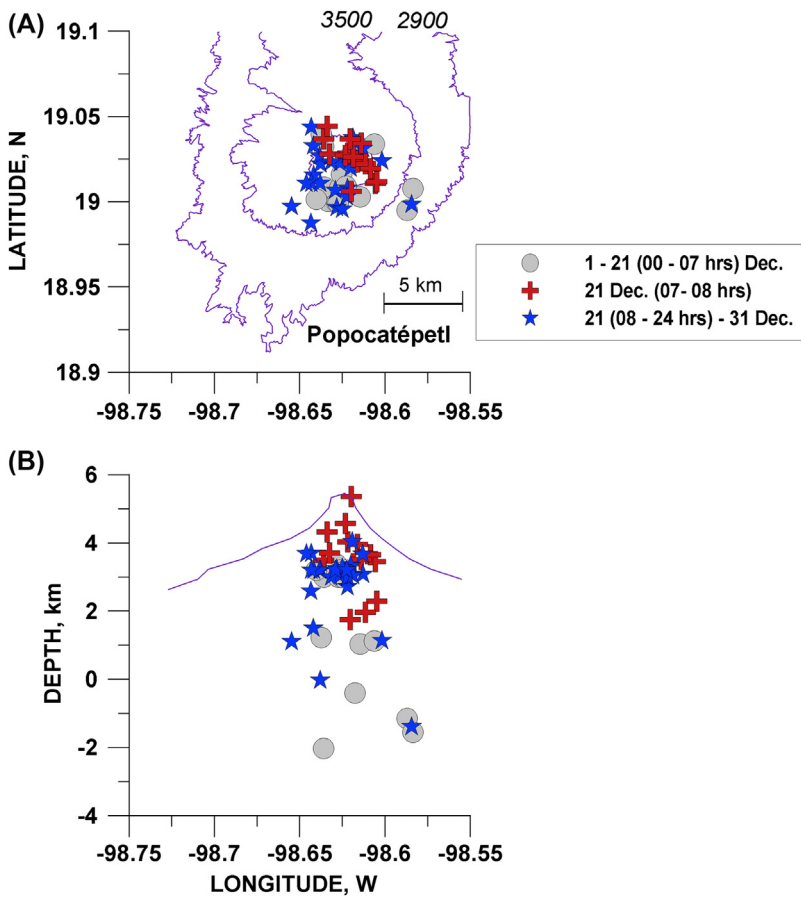


**FIGURE 6.11** Distribution of epicenters (A) and hypocenters (B) located at Popocatepetl volcano during December 1994. The contours of 3500 and 2900 m show the relief of Popocatepetl volcano. Data from the catalog of earthquakes prepared by J. Lermo-Samaniego (personal communication, 2003).



### 6.2.4 Soufrière Hills Volcano, Montserrat

Andesitic stratovolcano Soufrière Hills volcano (SHV) with a height of 915 m is situated in the southern part of Montserrat Island, West Indies. After about 370 years of repose (Simkin and Siebert, 1994), its eruption began on July 18, 1995, with the steam-and-ash venting from the north-west crater. The eruption activity passed through the phreatic and magmatic explosions, growth and collapse of lava domes accompanied by pyroclastic flows, and a great directed explosion. It continued for more than 15 years.

About 10,000 volcano-tectonic and hybrid earthquakes with magnitudes from  $-0.5$  to 2 were located from July 28, 1995, to the end of February 1997 with high precision (Aspinall et al., 1998; Miller et al., 1998). They followed all the main events in the volcanic activity. Earthquakes occurred within three epicentral zones: beneath SHV, along the structure extending to the NE of the volcano, and beneath St. George Hill (SGH), which is situated about 5 km to the NW of the volcano. All the events were recorded at shallow depths, mainly from 5 km b.s.l. to the surface. Fig. 6.12 shows the temporal variations in the depths of seismic events during 1995–97 for the

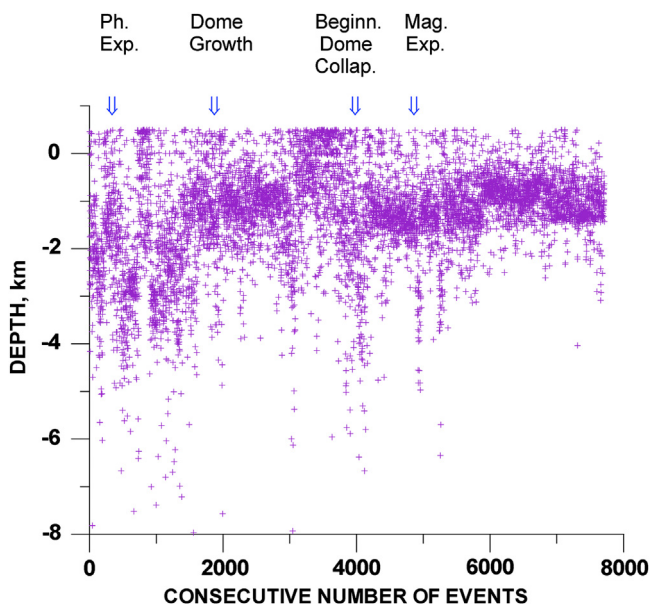


**FIGURE 6.12** Depth variations of earthquakes located beneath Soufrière Hills volcano from July 28, 1995 to February 28, 1997. Depth 0 km marks the sea level. Arrows show the main episodes in volcanic activity (Ph. Exp., phreatic explosion of August 21, 1995; Dome Growth and Beginn. Dome Collap., the initial growth and the beginning of dome collapses; Mag. Exp., magmatic explosion of September 17, 1996). Data from the catalog of the Montserrat Volcano Observatory.

epicentral zone of the volcano together with the main events in the volcanic activity. The tendency in shallowing of earthquake foci during the first year of activity from the phreatic explosion up to the beginning of the dome collapses is also shown in the figure.

We discuss here the variations in seismic activity preceding two important moments in the volcanic activity of SHV: the phreatic explosion of August 21, 1995, and the magmatic explosion of September 17, 1996.

#### 6.2.4.1 Phreatic Explosion of August 21, 1995

It was the first large phreatic explosion during the 1995–97 eruptive sequence. It blanketed the nearest city Plymouth, situated at a distance of about 6 km from the volcano, in a thick ash cloud and caused darkness for about 15 minutes (Chronology, 2002).

Figs. 6.13 and 6.14 show the temporal variations in earthquake number and the distribution of epicenters and hypocenters of earthquakes from August 1 to August 25. An increase in seismic activity began with a 2-day swarm of earthquakes along the NE extend of the SHV from August 5 to August 6, then a new 2-day swarm was recorded beneath SGH that

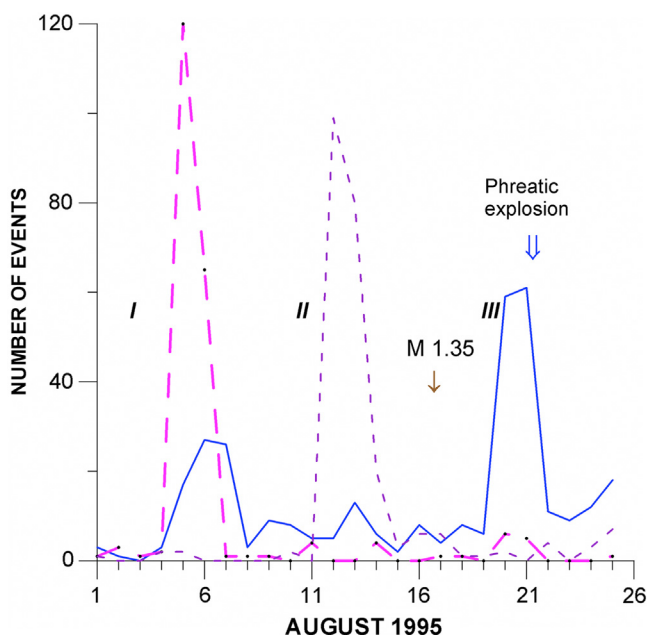


**FIGURE 6.13** Variations in daily earthquake numbers before the August 21, 1995 phreatic explosion at Soufrière Hills volcano. I, II, and III are the earthquake sequences that occurred within the NE extend from the volcano (I), beneath St. George's Hill (II), and beneath the volcano (III). The largest earthquake is shown by a *single arrow* and the phreatic explosion is shown by a *double arrow*. Data from the catalog of the Montserrat Volcano Observatory.

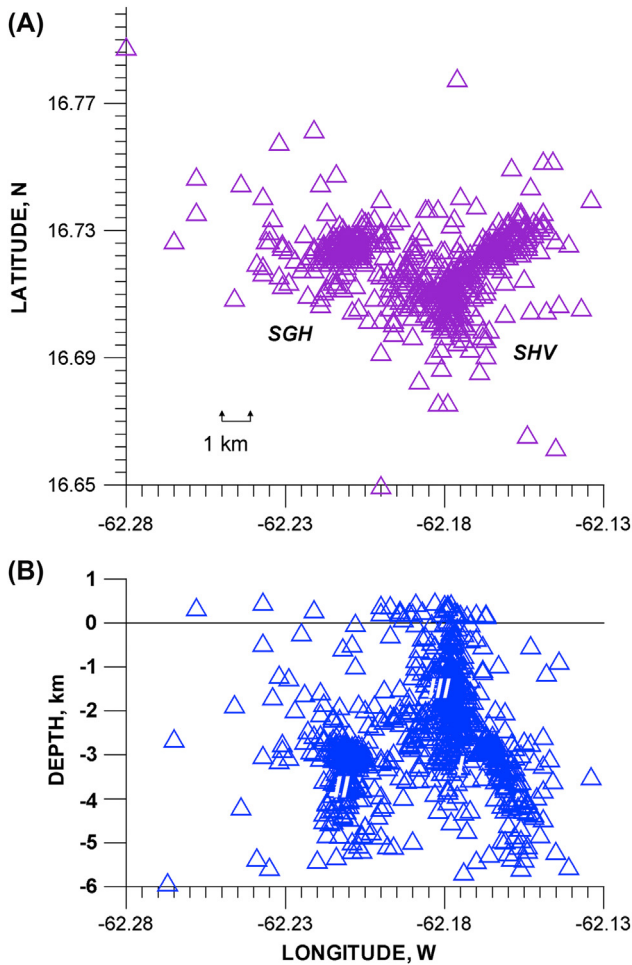


**FIGURE 6.14** Distribution of epicenters (A) and hypocenters (B) of earthquakes located before the August 21, 1995 phreatic explosion at Soufrière Hills volcano (SHV). *SGH* indicates St. George's Hill. I, II, and III in B are the earthquake sequences that occurred within the NE extend from the volcano (I), beneath St. George's Hill (II), and beneath the volcano (III). The zero-line of sea-level depth is shown. Data from the catalog of the Montserrat Volcano Observatory.

continued from August 12 to August 13. The seismicity beneath the volcano was low up to August 20. During 2 days, that is August 20 and August 21, about 140 earthquakes were recorded. The explosion occurred at 08:03 a.m. (LT) of August 21 (B. Voight, personal communication, 2002) during the maximum level of seismic activity below the volcano. Just after the explosion, the number of seismic events sharply decreased.

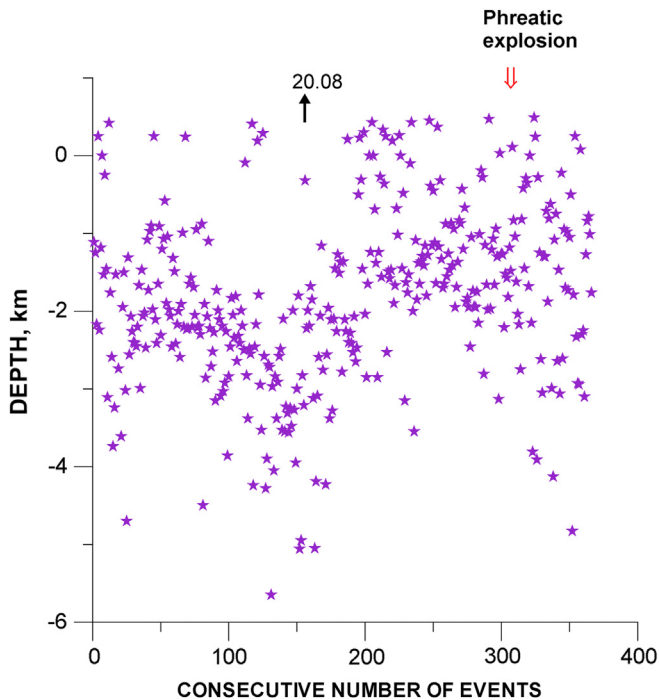


**FIGURE 6.15** Temporal variations in earthquake depths beneath the volcano before the August 21, 1995 phreatic explosion at Soufrière Hills volcano. The phreatic explosion is shown by double arrow. The beginning of the August 20–21 swarm is shown by a single arrow. Data from the catalog of the Montserrat Volcano Observatory.

The seismic events recorded beneath the volcano were clustered within a vertical column-shaped volume extending from 3 km b.s.l. to the summit of the volcano. The foci tended to be shallower before the phreatic explosion. Fig. 6.15 shows that the focal depths of earthquakes changed from 1 to 3 km b.s.l. on August 1–19 to the depth range from 2 km b.s.l. to the summit of the volcano on August 20–21. The events associated with the extension of the NE structure of the volcano, as well as the SGH earthquakes, were deeper and located at depths from 5 to 3 km b.s.l.

#### 6.2.4.2 Magmatic Explosion of September 17, 1996

A series of dome collapses, which began in July 1996 (see Fig. 6.12), led to the September 17 (23:46 LT) magmatic explosion. Rocks and pumice fell in the south of the island, and houses were destroyed in the nearest village. A major ash plume rose to about 12 km, and about 600,000 tons of ash were deposited in southern Montserrat (Chronology, 2002).

The seismic activity that preceded the September 17 explosion is shown in Figs. 6.16–6.18. The earthquake foci, located during about 40 days of activity, are clustered mainly beneath the SHV forming a vertical column rising from a depth of 3 km b.s.l. to the summit of the volcano. The zones that were active before the August 21, 1995 phreatic explosion are now calm.

The swarm of earthquakes had no clear beginning; it began in the background of continuous seismic activity due to the process of partial collapsing of the dome that began in July 1996. Nevertheless, we can mark the gradual increase in earthquake numbers beginning from August 29; therefore, the swarm continued for about 23–24 days. All the earthquakes were of low magnitude; the largest event with a magnitude of 1.9 occurred just before the explosion of September 17. The explosion occurred during the apparent decrease in the number of seismic events; after the explosion, the number of earthquakes sharply decreased.

Fig. 6.18 shows that the seismic process developed permanently at depths between 1 and 2 km. No focal depth migration to the surface was noted before the explosion.

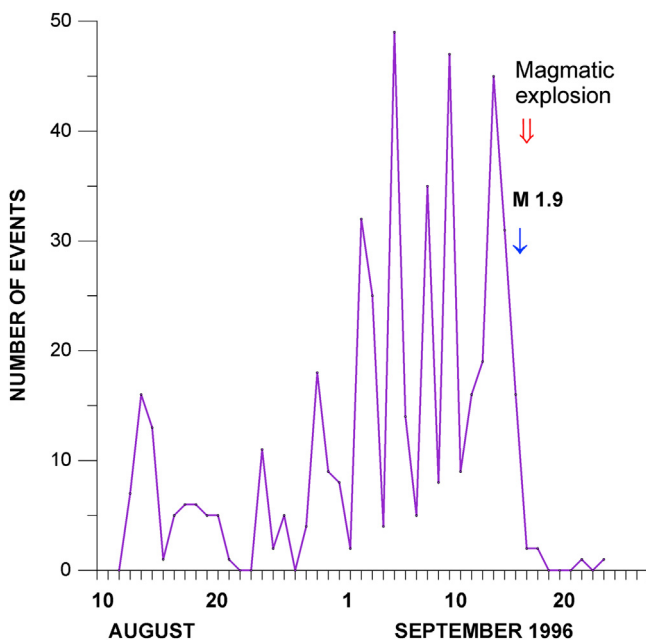


**FIGURE 6.16** Variations in daily earthquake numbers before the September 17, 1996 magmatic explosion at Soufrière Hills volcano. The appearance of the largest earthquake is shown by a *single arrow* and the beginning of the magmatic explosion is shown by a *double arrow*. Data from the catalog of the Montserrat Volcano Observatory.

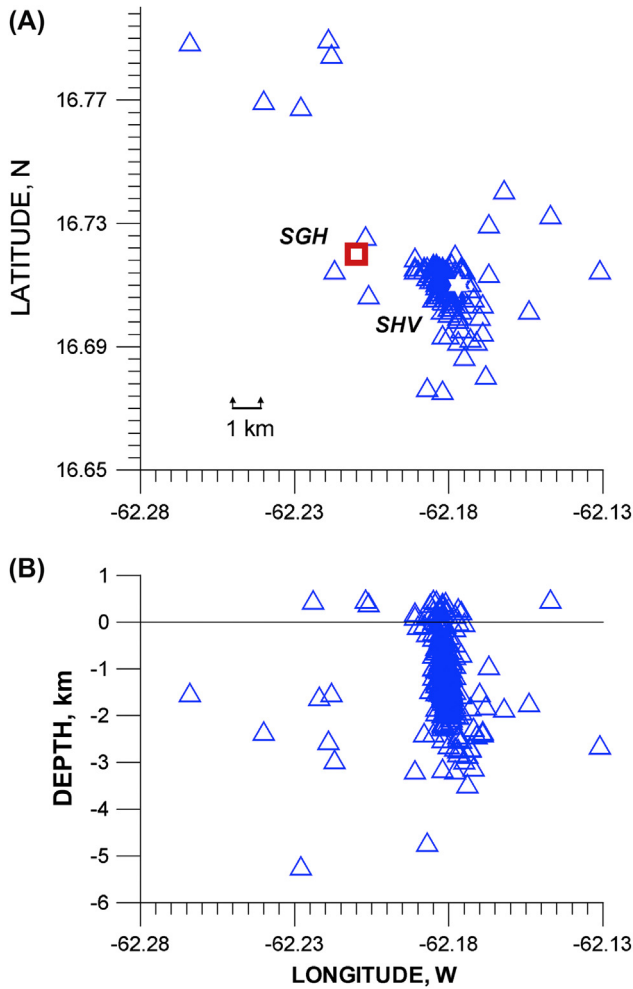


**FIGURE 6.17** Distribution of epicenters (A) and hypocenters (B) of earthquakes located before the September 17, 1996 magmatic explosion at Soufrière Hills volcano (SHV). The *square* marks St. George's Hill (SGH), and the *star* marks SHV. Data from the catalog of the Montserrat Volcano Observatory.

## 6.3 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH LAVA EXTRUSIONS

### 6.3.1 Volcán De Colima, Mexico

The recent 1997–2017 unrest at Volcán de Colima was characterized by the building destruction of seven andesitic block-lava domes. The first lava dome began to grow up in the summit crater on November 20, 1998.

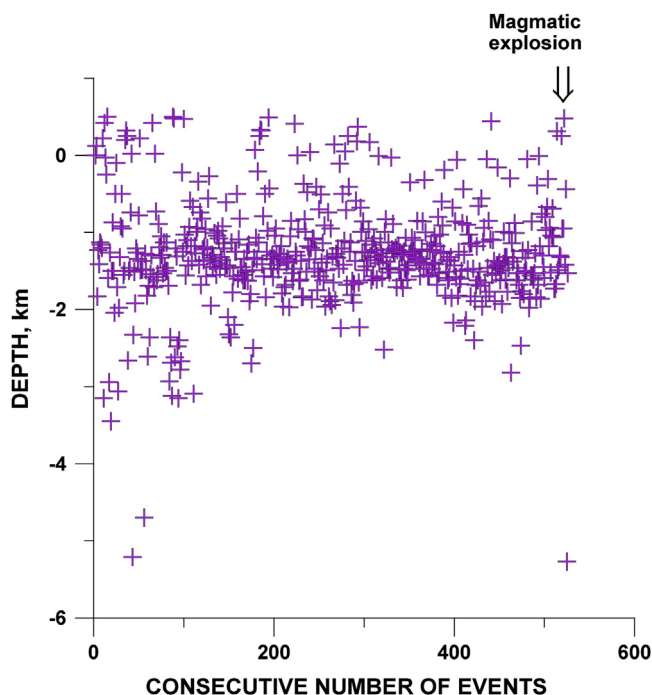


**FIGURE 6.18** Temporal variations in earthquake depths beneath the volcano before the September 17, 1996 magmatic explosion at Soufrière Hills volcano. The magmatic explosion is shown by a double arrow. Data from the catalog of the Montserrat Volcano Observatory.

The dome quickly spilled over the SW crater rim and fed block-lava flows that descended in three branches on the southern slope of the volcano, to the maximum distance from the crater of 3.8 km; the advance of the block-lava flows was accompanied by pyroclastic flows (Zobin et al., 2002a,b).

Fig. 6.19 shows the temporal development of seismic activity before the 1998 eruption. The seismic activity began on November 28, 1997, and continued till November 20, 1998. About 15,000 small volcano-tectonic earthquakes were recorded. The majority of them were clustered within five earthquake swarms. The lava extrusion began when the last, fifth, swarm reached its maximum in the number of earthquakes.

The distribution of epicenters and hypocenters of earthquakes with magnitude  $M_w$  from  $-0.5$  to  $2.7$  recorded from November 1997 to November 1998 is shown in Fig. 6.20. Only the earthquakes located within the area of the seismic network are plotted, as those outside the network had larger location uncertainties. The epicenters were located within the Colima volcanic complex and covered an area of about  $50 \text{ km}^2$ . The majority of events clustered in the area of the Colima volcano's active crater and between the volcanic edifices of Colima and Nevado volcanoes.

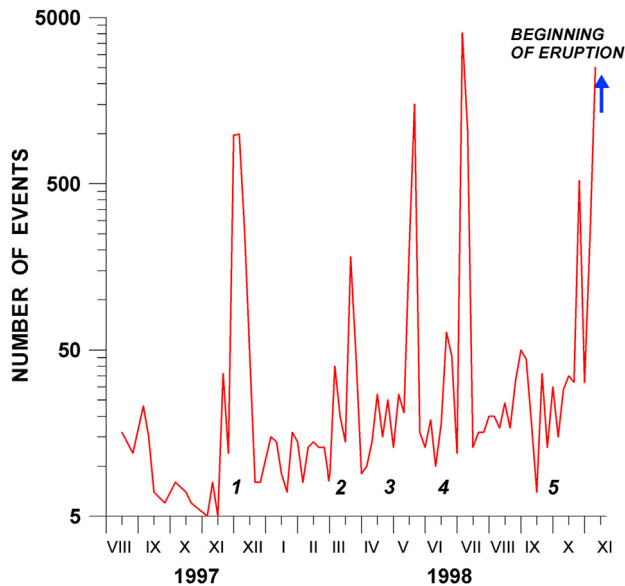


**FIGURE 6.19** Variations in the numbers of earthquakes recorded during 1997–98 before the November 20, 1998 lava extrusion at Volcán de Colima. Counted events are those recorded by a minimum four stations over 5-day intervals. Numbers 1–5 identify individual swarms. *Data from the catalog of Colima Volcano Observatory.*

The majority of hypocenters were located at depths ranging from 10 km (mainly from 5 km) b.s.l. to the crater zone of the volcanoes, or up to heights of about 4000 m. A dense clustering of hypocenters was observed within a narrow depth interval of 1–4 km a.s.l. within the volcanic constructions. Fig. 6.20B shows a space free of earthquake foci just below the volcano at the depths of 0–10 km b.s.l. This free space could correspond to a zone of possible magma storage.

Fig. 6.21 shows the depth variations of earthquakes in time. It is difficult to note any variations in depth for the events within the individual swarms. However, there were substantial variations in foci depth from swarm to swarm. It is possible to distinguish two stages in depth variations. The first stage includes the first three swarms from November 1997 to May 1998; the depths gradually decreased from swarm to swarm. During June–July (Swarm 4), the depths again increased, with their distribution similar to the depth distribution of the first November 1997–December 1997 swarm. The last swarm (Swarm 5) that occurred just before the eruption was shallower, with events occurring only from 1 to 4 km a.s.l.

The eruption of Volcán de Colima continued till February 2017. New six lava domes were formed and destructed but no more swarms of volcano-tectonic earthquakes before the lava extrusions were recorded.



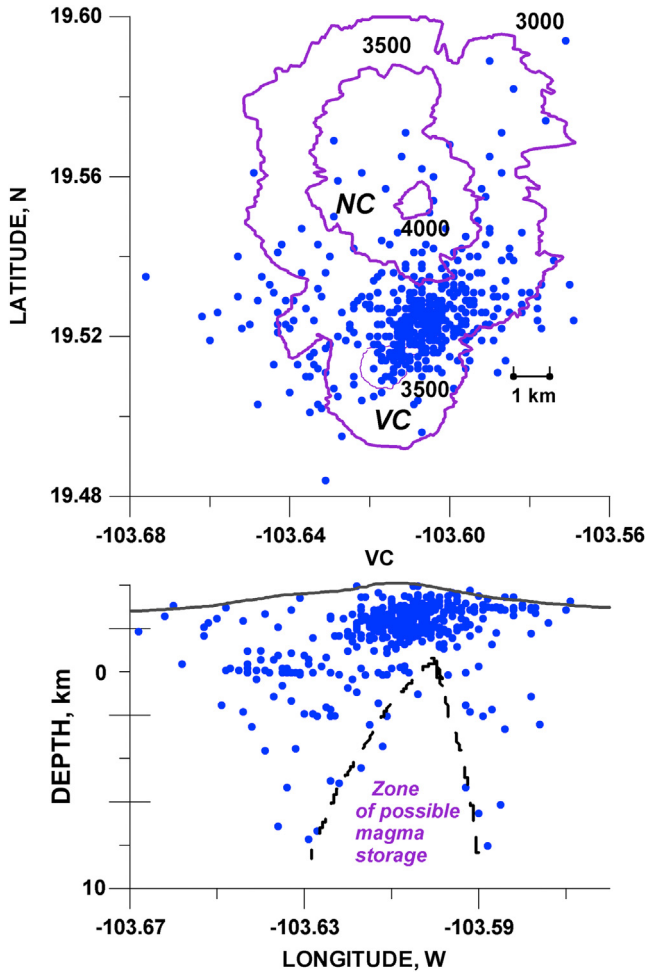


**FIGURE 6.20** The distribution of epicenters (above) and hypocenters (below) of earthquakes located during the seismic crisis of 1997–98 at Volcán de Colima. The contours of 3000, 3500, and 4000 m show the relief of Colima volcanic complex. VC, Volcán de Colima; NC, Nevado de Colima. An earthquake-free zone is outlined in the cross-section. From Zobin, V.M., González Amezcua, M., Reyes-Dávila, G.A., Domínguez, T., Cerda Chacón J.C., and Chávez Álvarez, J.M., 2002. The comparative characteristics of the 1997–1998 seismic swarms preceding the November 1998 eruption of Volcán de Colima, Mexico. *Journal of Volcanology and Geothermal Research*. 117, 47–60.

### 6.3.2 La Soufriere volcano, St Vincent Island

The La Soufrière volcano is located in the northern part of the St Vincent Island, the Lesser Antilles, situated in the Caribbean Sea. It is a 1220-m-high basaltic-andesitic stratovolcano with a summit crater  $\sim 1.6$  km in diameter and 300–600 m deep. The 2020–21 eruption consisted of two distinct

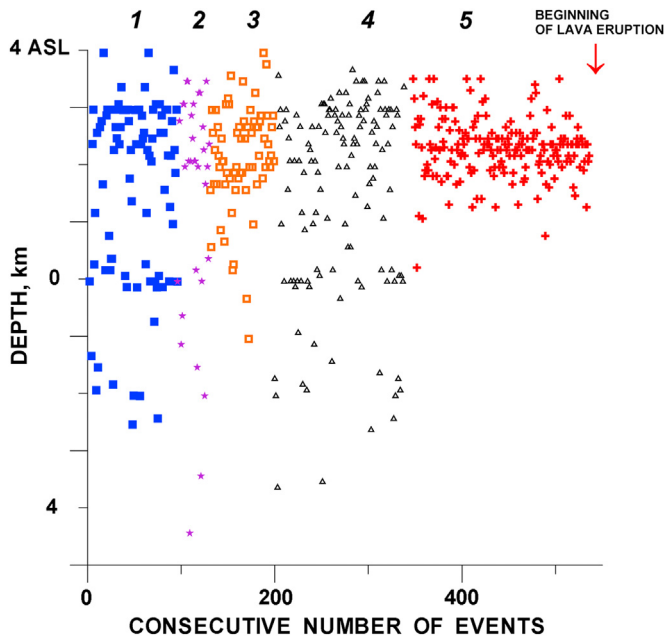


**FIGURE 6.21** Temporal variations of the depths of earthquakes located during the five swarms (numbers 1–5) at Volcán de Colima. Events in each swarm are marked by particular symbols. From Zobin, V.M., González Amezcua, M., Reyes-Dávila, G.A., Domínguez, T., Cerda Chacón J. C., and Chávez Álvarez, J.M., 2002. The comparative characteristics of the 1997–1998 seismic swarms preceding the November 1998 eruption of Volcán de Colima, Mexico. *Journal of Volcanology and Geothermal Research*. 117, 47–60.

phases. The initial effusive phase began on December 27, 2020 and was characterized by the extrusion of a small lava dome inside the south-west sector of the summit crater (Fig. 6.22A). Effusive activity continued for three months with a rapid increase in effusion rate in early April 2021. It was followed by an explosive phase between April 9 and April 25, 2021, which consisted of more than 30 discrete sub-Plinian and Vulcanian explosions (Fig. 6.22B) (Joseph et al., 2022; Stinton, 2023).

The seismic activity began to accompany the lava dome extrusion only beginning on January 6, 2021 (Joseph et al., 2022). The seismic catalog of local events was collected by the Seismic Research Centre at The University of the West Indies using up to seven seismic stations on St. Vincent. The seismicity stayed at a low level (an average of two events per day) up to January 17, when the first notable increase in seismic activity was detected with 60 events recorded. After January 17, seismicity increased steadily, at a rate of approximately four events per day. However, at the end of March, seismicity decreased before increasing significantly again (at a rate of  $\sim 30$  events/day) during the buildup to the start of the explosive phase (Thompson et al., 2022). This sharp decrease in seismic activity continued for about 10 days (Fig. 6.23).

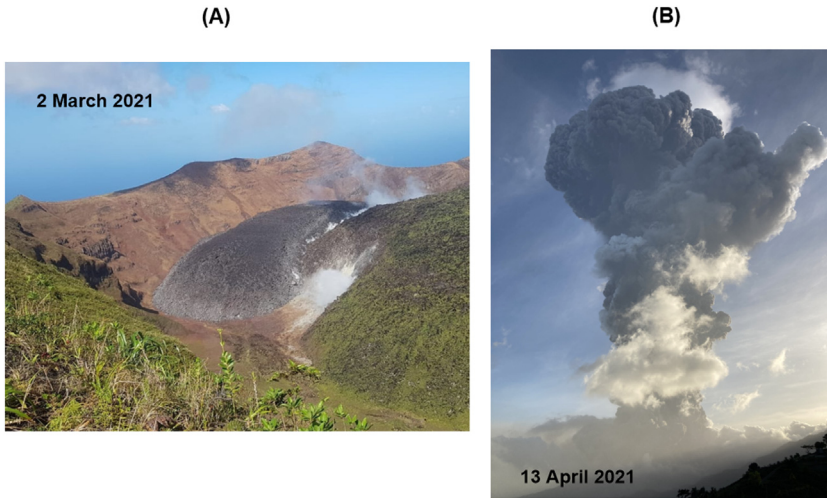


**FIGURE 6.22** Growing lava dome (A) and one of Vulcanian explosions that destroyed the dome (B) during the 2020–21 eruption at La Soufrière volcano. (A) From Stinton, A., 2023. *Growth and evolution of the lava dome and coulée during the 2020–21 eruption of La Soufrière, St Vincent*. Geological Society, London, Special Publications, 539. (B) From Sparks, S.R.J., Aspinall, W.P., Barclay, J., Renfrew, I.A., Contreras-Arratia, R. and Stewart, R., 2023. *Analysis of magma flux and eruption intensity during the 2021 explosive activity at the Soufrière of St Vincent, West Indies*. Geological Society, London, Special Publications, 539.

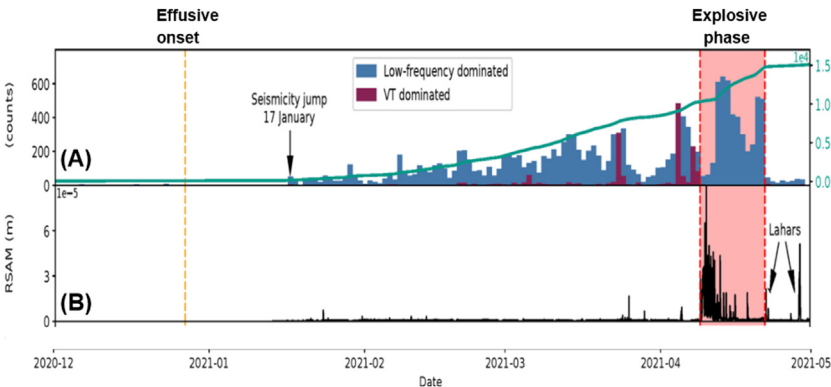


**FIGURE 6.23** Seismic monitoring data for the 2020–21 eruption at La Soufrière volcano (A). Daily (bars) and cumulative (teal line) seismicity observed and a marker on the first important increase (January 17). The bars in blue represent seismic events related to fluid dynamics (low-frequency and dome emplacement) and the bars in dark red represent VT events. (B). Real-time seismic-amplitude measurement (RSAM) values: calculated with 1 min windows and no overlap; they correlate with volcano-tectonic swarms and explosive phase. Modified from Joseph, E.P., Camejo, M., Cristopher, T. et al., 2022. *Responding to eruptive transitions during the 2020–2021 eruption of La Soufrière volcano, St. Vincent*. Nature Communications, 13, 4129, <https://doi.org/10.1038/s41467-022-31901-4>.

The seismic activity during the effusive phase of the eruption was represented mainly by low-magnitude volcano-tectonic and low-frequency earthquakes. The volcano-tectonic seismic swarms were recorded by most of the stations while most superficial low-frequency seismicity could only be observed in the 1–2 closest stations because of small amplitudes. During the explosive phase, only low-frequency earthquakes, accompanied by volcanic tremors, were recorded (Fig. 6.23) (Joseph et al., 2022; Thompson et al., 2022).

Volcano-tectonic earthquakes of the effusive phase were represented mainly in two swarms (Fig. 6.24). The first of them occurred during March 23–24, 2021 (226 events) with >95% located at depths shallower than 5 km. The swarm, recorded April 5–6, 2021 (476 events), was characterized by an abrupt transition to deeper locations. These two swarms of volcano-tectonic earthquakes preceded and accompanied a

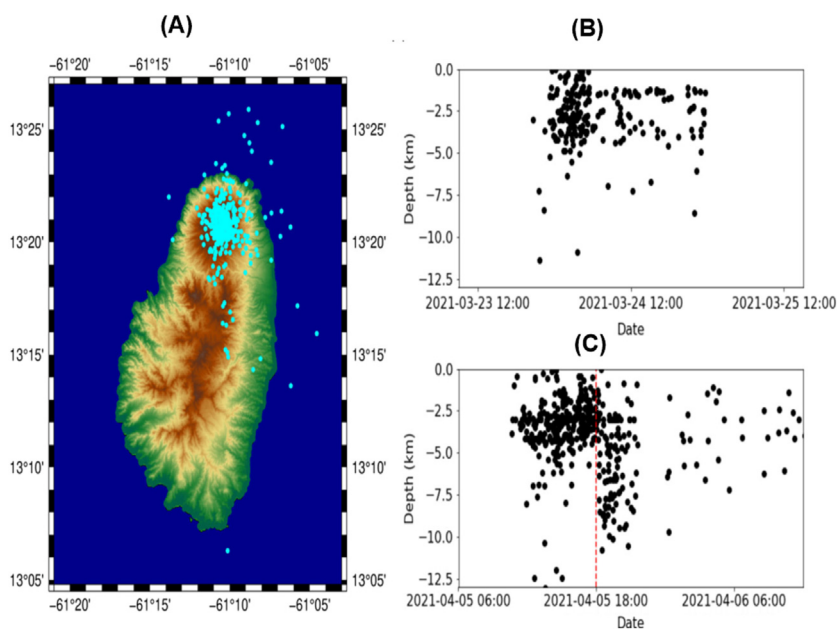


**FIGURE 6.24** Seismicity located during March 23–24 and April 5–6 volcano-tectonic swarms. (A) Epicenters calculated showing concentric distribution. (B) Temporal evolution of depths during March 23–24, most seismicity is shallower than 5 km. (C) Temporal evolution of depths during April 5–6, seismicity is shallower than 5 km until 18:00 UTC on April 5, when locations transitioned to deeper levels. This moment is indicated with vertical line. From Joseph, E.P., Camejo, M., Christopher, T. et al., 2022. Responding to eruptive transitions during the 2020–2021 eruption of La Soufrière volcano, St. Vincent. *Nature Communications*, 13, 4129, <https://doi.org/10.1038/s41467-022-31901-4>.

sharp decrease in seismic activity mentioned above. The deeper location of the volcano-tectonic sources three days before the final stage in the lava dome construction and the beginning of the explosive phase may be interpreted because of a new additional volume of magma ascending from  $\sim 10$  km depth. As is seen in Fig. 6.24, volcano-tectonic earthquakes dominated till the beginning of the explosive phase (Joseph et al., 2022).

## 6.4 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH FLANK ERUPTIONS

### 6.4.1 Sakurajima Volcano, Japan 6.3.2 La Soufriere

Sakurajima volcano is an andesitic stratovolcano (altitude of 1110 m) that is situated in the southern part of the Aira caldera, Southern Kyushu. The volcano had repeated flank and summit eruptions during historic times (Ishihara, 1990). The 1914 eruption of the volcano was one of the largest volcanic events in the 20th century in Japan, and it was one of the first volcanic eruptions that were studied using instrumental seismic observations (Omori, 1914; Yokoyama, 1997).

The eruption began on January 12, 1914, at around 08:00 hours (LT) with the emissions of white smoke from the summit crater, and 2 hours later explosions occurred at the western and eastern slopes, where the vents were opened. Just after midnight of January 13, the strongest explosions were observed and at about 20 hours lava began to effuse from the craterlets on both the eastern and western slopes. The eruption continued till May 1915 (Yokoyama, 1997).

Seismic activity near the volcano was recorded in Kagoshima by the local seismic station beginning on January 11 (Omori, 1914). The recording drum of the low-sensitivity seismographs (magnification  $\times 5$ – $10$ ) was triggered into motion by intense shocks over a certain threshold (Fig. 6.25). Their magnitudes were 4.8–5.2, according to Yokoyama (1997). The higher sensitivity seismic station in Nagasaki (epicentral distance was about 148 km, but the magnifications were of  $\times 20$  and  $\times 120$ ) began to record the swarm at the lower threshold on January 9. Earthquakes increased in number and magnitude, with the maximum activity at midnight on January 12. After the opening of flank vents in the morning of January, earthquake activity rapidly decreased. Then, at 18:30 on January 12, with a background of seismic calmness, the magnitude  $M_w$  7.0 earthquake (Zobin, 2001) shock occurred. About 26 hours later, the great lava flank extrusion began. About  $1.6 \text{ km}^3$  of lavas overflowed from the volcano, which connected the volcanic island with Oosumi Peninsula (Yokoyama, 1997).

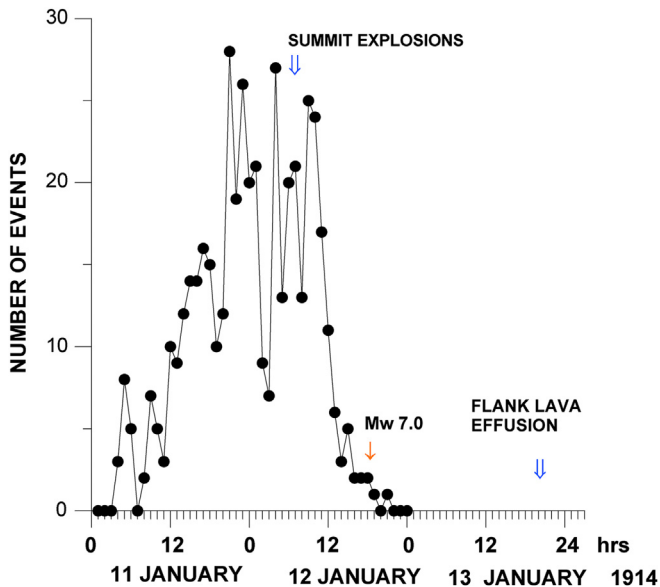


**FIGURE 6.25** Temporal variations of hourly number of volcano-tectonic earthquakes at Sakurajima volcano recorded by the Gray–Milne–Ewing seismograph in Kagoshima at a distance of 10 km from eruptive centers. A single arrow shows the largest earthquake, while double arrows show the eruptive events. From Omori, F., 1914. *The Sakura-jima eruptions and earthquakes, I. Bulletin of the Imperial Earthquake Investigation Committee* 8, 5–34.

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## Chapter 7

# Volcano-Tectonic Earthquakes at Dacitic Volcanoes: Case Studies

Eruptions of dacitic volcanoes may occur from the summit crater and from the new vents on the flank of a volcano. The high viscosity and relatively low-temperature magmas with great amounts of volatiles are characterized by low mobility and high explosivity. The magma may construct a lava plug dome, cryptodome, or lava spine in the crater of the volcano. Large volumes of volatiles can produce numerous explosions of different intensity. Volcano-tectonic earthquakes can be observed during the movement of magma to the surface before lava extrusion. Lava effusion from the summit and flank vents may be accompanied by volcanic tremors. Dacitic volcanoes have low altitude and stratovolcano shape.

### 7.1 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH SUMMIT ERUPTIONS

Summit eruption of dacitic volcanoes may consist of the growth of a dome in the crater with the formation of lava flows and/or a great explosion or a sequence of numerous small explosions. The most famous summit eruptions at dacitic volcanoes occurred at Mount St. Helens (MSH) in the United States, Usu in Japan, and Pinatubo in the Philippines. Volcano-tectonic earthquakes are related to lava dome eruptions and magmatic explosions.

#### 7.1.1 Mount St. Helens, Cascades

MSH is a stratovolcano of 2950 m in height. It is situated within the Cascade Range in the western state of Washington, United States. The eruptive history of MSH began about 40,000 years ago with dacitic volcanism. The products of eruptive activity during the past 2500 years were represented mainly by dacite–andesite–dacites. The volcano had been apparently dormant since 1857 (Mullineaux and Crandell, 1981).

##### 7.1.1.1 Eruption of 1980

The first 1980 steam-blast eruption from its summit crater was observed on March 27 and was followed by a sequence of small explosive eruptions up

to the climactic eruption on the morning of May 18. This explosive eruption of dacitic magma formed a vertical Plinian column with more than 25 km high. The northward-directed lateral blast of the bulging north flank of the volcano erupted about  $2 \text{ km}^3$  of cold and juvenile material and devastated an area of nearly  $600 \text{ km}^2$ . A new crater, with an area of  $1.5 \text{ km} \times 3.0 \text{ km}$  and a depth of about 500 m, was formed on the northern flank of the volcano. The height of the volcano was reduced by about 400 m. Since then, several smaller magmatic eruptions occurred during 1980 producing pyroclastic flows, ash falls, and several lava domes (Christiansen and Peterson, 1981). Lava domes began to grow in the crater beginning in June 1980 (Moore et al., 1981). The eruption terminated in 1986.

The seismic activity of MSH was monitored by the regional Pacific Northwest Seismic Network (PNSN). Fig. 7.1 shows the temporal variations in the daily number of events and their maximum magnitudes during three stages of activity: preeruptive, before the great May 18 explosion, and after the great May 18 explosion up to the end of August 1980. The 1980 swarm of volcano-tectonic shallow earthquakes began on March 16 with small events, and the first significant  $M_L$  4.2 event was recorded beneath MSH on March 20. The number and magnitude of earthquakes sharply increased starting March 25 with a peak number on March 27, when the first explosion occurred. After the beginning of the eruption, the number of earthquakes gradually decreased for 1 month; then their number was more or less stable up to May 18, when a cataclysmic explosion occurred. The largest earthquake with  $M_w$  5.7 (Zobin, 2001) occurred on May 18 in the background of low-level seismic activity, practically simultaneously with the great directed blast of the volcano.

Fig. 7.2 shows the location of earthquakes that occurred during March 1980–May 1980. Before the March 27 explosions, a total of 93 earthquakes were located beneath MSH and to its SE. The seismic events formed two clusters: (1) shallow, at the depth range of 0–5 km, and (2) deep cluster, at the depth range from 18 to 25 km. During the active period from March 27 to May 18, a total of 627 located earthquakes occurred beneath the volcanic edifice, mainly at the depth range from 0 to 10 km; the deep cluster disappeared. The largest earthquake of May 18 was located at a depth of 2.8 km beneath the NW flank of MSH.

After the great May 18 explosion, seismic activity beneath the crater continued and was associated with a few magmatic eruptions of the volcano from May 1980 to August 1980 (Weaver et al., 1981; Malone et al., 1981). At the same time, the regions to the north and to the south of the volcano became seismically active (Fig. 7.3). The 50-km-length submeridional linear structure with the volcano in its middle section was activated. Seismic foci to the south of the volcano clustered at a depth range from 20 to 1 km; the northern foci formed a cluster with a length of about 15 km and within the same depth range from the surface to 20 km.

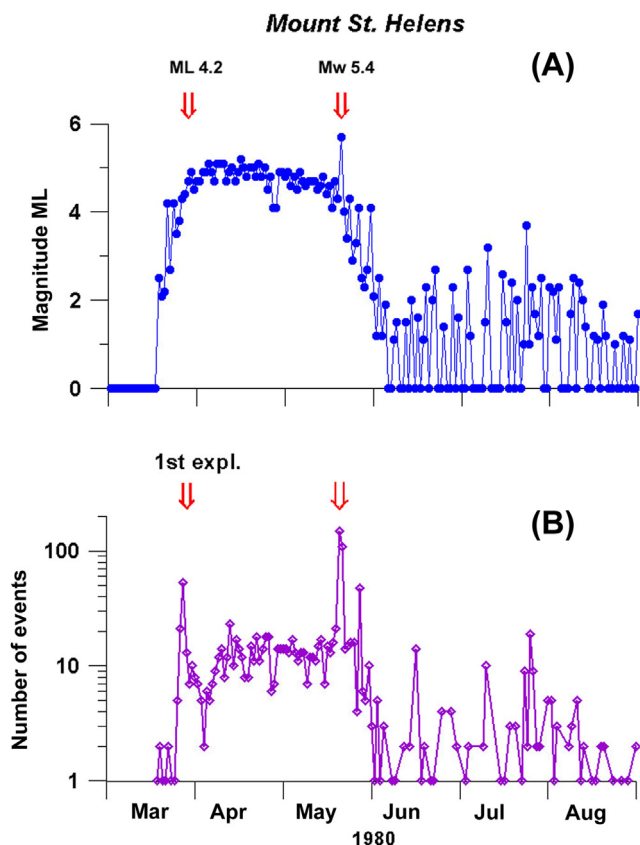


**FIGURE 7.1** Variations in daily maximum magnitude (A) and the number of earthquakes (B) located during the March–August 1980 seismo-volcanic activity at Mount St. Helens (MSH). The double arrows show eruptive events. Data from the Pacific Northwest Seismic Network (PNSN) catalog.

The seismic foci just beneath the volcano occupied space between deep (16–22 km) and shallow (from a depth of 10 km to the surface) groups of earthquakes recorded before the May 18 explosion.

During the period between 1980 and 1986, approximately 20 small eruptions were recorded. Then the active period changed with the quiescence between 1987 and 2004. Fig. 7.4 shows the change in the seismic activity in late 1987, when the first of several years-long swarms of “deep” (> 3 km) volcano-tectonic events began. The depth of these swarms contrasted sharply with the exclusively shallow (< 3 km) precursory swarms associated with 19 of the 20 following May 18, 1980 dome-building eruptions. In September 2004, a new large eruption began.

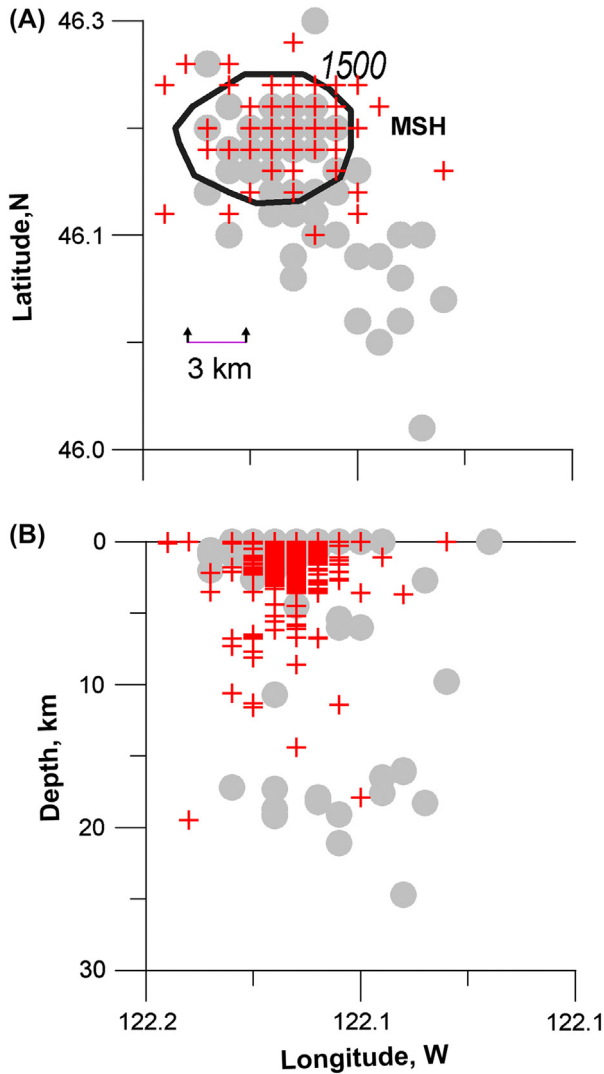


**FIGURE 7.2** Epicenters (A) and hypocenters (B) of earthquakes located during March 1980 to May 1980 at Mount St. Helens (MSH). The earthquakes located before March 27, when the first explosion occurred, are shown by *circles*; the earthquakes located from March 27 to May 18 are shown by *crosses*. The position of volcano MSH is shown by the 1500 m contour. Data from the Pacific Northwest Seismic Network (PNSN) catalog.

## 7.1.2 Usu Volcano, Hokkaido

### 7.1.2.1 Eruption of 1977–78

The eruption of 1977–78 occurred in the southeastern part of the summit crater (Fig. 7.5). Four new craters were formed during a sequence of

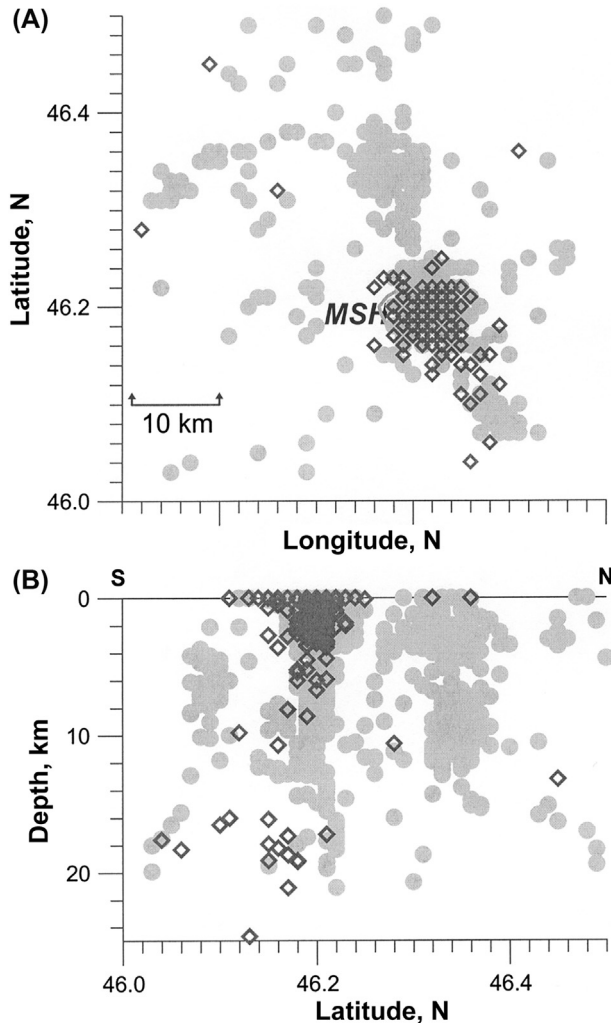


**FIGURE 7.3** Following May 18, 1980 seismic activity in the region of Mount St. Helens (MSH) observed from May 19, 1980, to August 31, 1981. Epicenters (A) and hypocenters (B) of earthquakes (N–S cross-section). The earthquakes, recorded before the May 18 explosion, are shown by diamonds; the earthquakes, recorded after the May 18 explosion, are shown by circles. Data from the Pacific Northwest Seismic Network (PNSN) catalog.

magmatic explosions on August 7–August 14. The volume of ejected pumice material was about  $0.01 \text{ km}^3$  (Yokoyama et al., 1981).

The eruption was preceded by 32 hours of seismic activity (Fig. 7.6). The earthquake swarm consisted of about 3000 small (magnitude 1–2) volcano-tectonic earthquakes that were located beneath the summit crater at a depth from 0 to 2 km (Fig. 7.5). The number of earthquakes gradually increased

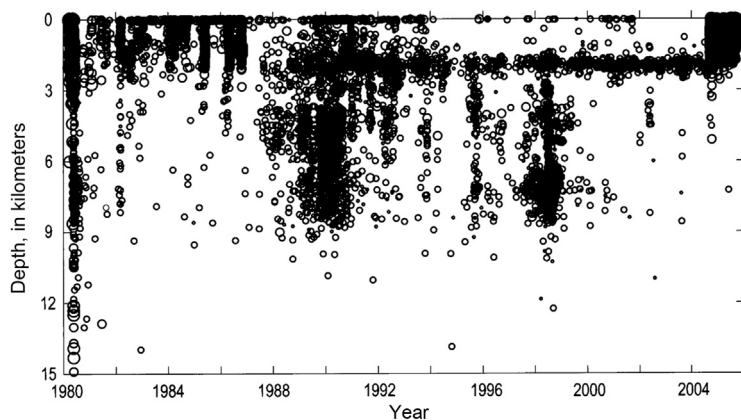


**FIGURE 7.4** Time versus depth plot showing all earthquakes (circles) with well-constrained locations occurring at Mount St. Helens (MSH) between 1980 and 2005. Depths are relative to a datum corresponding to an average altitude of stations in the MSH network, roughly 1.1 km a.s.l. From Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Prof. Paper 1750, Washington, pp. 27–60.

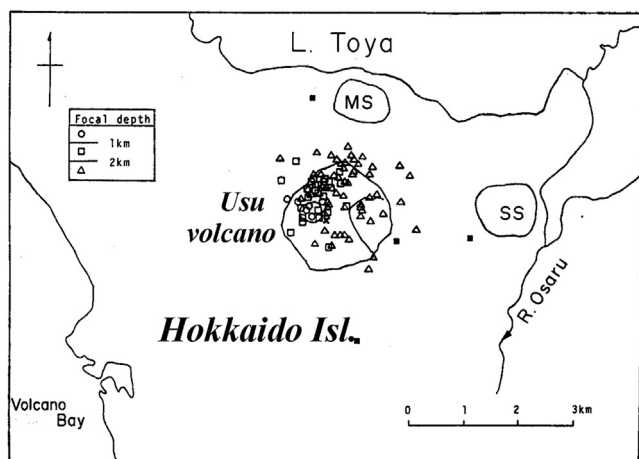


**FIGURE 7.5** Epicenters of earthquakes located just before the 7 August 1977 eruption of Usu volcano. MS is the 1910 cryptodome Meiji-shinzan; SS is the 1943–45 lava dome Showa-shinzan. Solid squares show the position of seismometers; the cross indicates the position of crater 1. From Yokoyama, I., Yamashita, H., Watanabe, H., Okada, H., 1981. Geophysical characteristics of dacite volcanism – the 1977–1978 eruption of Usu volcano. *Journal of Volcanology and Geothermal Research* 9, 335–358.

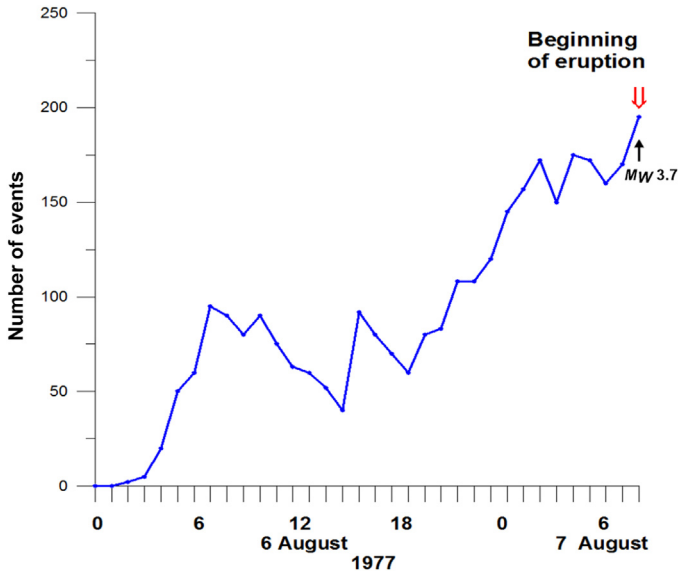


**FIGURE 7.6** Temporal variations in hourly number of earthquakes recorded at the station JMA-A located on the somma of Usu 32 hours before the 1977 eruption. The *double arrow* shows the beginning of the eruption; the *single arrow* shows the maximum magnitude earthquake. From Yokoyama, I., Yamashita, H., Watanabe, H., Okada, H., 1981. Geophysical characteristics of dacite volcanism – the 1977–1978 eruption of Usu volcano. *Journal of Volcanology and Geothermal Research* 9, 335–358.

with time, and the eruption began when the frequency of events reached its maximum. The maximum magnitude  $M_w$  3.7 earthquake was observed half an hour before the beginning of the eruption (Yokoyama and Seino, 2000; Yokoyama, 2001). The new eruptive craters were situated within the epicentral zone of earthquakes. The alarm was issued by the Japan Meteorological Agency (JMA), and there were no victims (Yokoyama et al., 1981).

During the August 8–August 14 stage of pumice explosions, strong explosions were preceded and followed by a sharp decrease in seismic activity (Fig. 7.7). The stage of calm after the explosions continued for 3–10 hours, and then seismic activity gradually increased again (Suzuki and Kasahara, 1979).

### 7.1.3 Unzen Volcano, Kyushu

Unzen volcano is a composite volcano of 1359 m in height and consists of multiple volcanic cones of dacite to andesite. Following 198 years of dormancy, a small phreatic eruption started at its summit on November 17, 1990. During February 1991–April 1991, a sequence of phreatomagmatic eruptions followed a resumed phreatic eruption on February 12; they

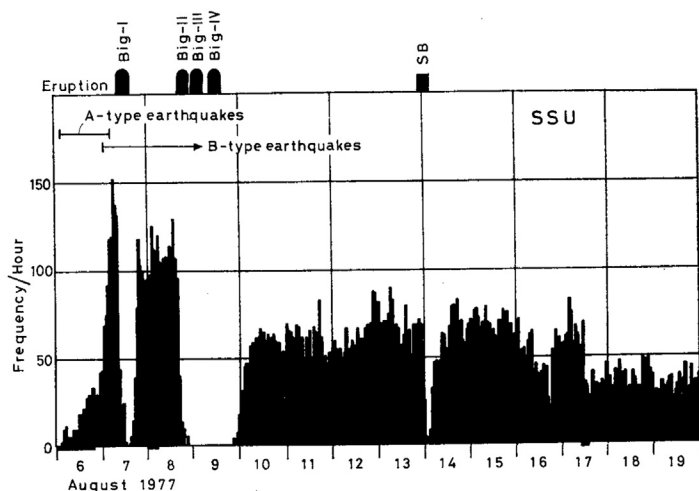


**FIGURE 7.7** Variations of hourly earthquake frequency at the station SSU situated at a distance of 2.5 km east from the crater. ( $M > 0.8$ ) from 6 to 19 August 1977. It is clearly seen that there is an absence of seismic events just after the large explosions marked as Big-I, II, III, IV and SB. From Suzuki, S., and Kasahara, M., 1979. *Seismic activity immediately before and in the early stage of the 1977 eruption of Usu volcano, Hokkaido, Japan. J. Fac. Sci., Hokk. Univ., ser. VII 6, 239–254.*

developed into a dacitic dome eruption on May 20, 1991. The growth of the dome continued for about 4 years accompanied by repeating partial collapses and pyroclastic flows. The total volume of erupted materials was  $2.1 \times 10^8 \text{ m}^3$  (Nakada et al., 1999).

Seismic activity began 1 year before the phreatic eruption (Nakada et al., 1999; Umakoshi et al., 2001). Fig. 7.8 shows the temporal variations in the number of earthquakes. Five swarms of volcano-tectonic earthquakes were recorded up to the moment of lava extrusion. The first two swarms were not followed by volcanic activity. Swarm 1 occurred at a distance of about 20 km to the west of the volcano at depths from 10 to 20 km below the ancient Chijiwa caldera (Fig. 7.9A). It began on November 21, 1989, and continued for 4 days. About 600 earthquakes were located with the maximum  $M$  3.3 event. Swarm 2, which occurred from July 1990 to August 1990, was shallower, and it was located within the depth range from 15 to 5 km. The largest earthquake with a magnitude of 3.9 was observed on July 7. This swarm occupied the territory between the first swarm epicenters and the volcano (Fig. 7.9A).

The next three impulses of seismic activity occurred in November 1990 and during February to March 1991 and May 1991. They were associated with eruptive events. Their hypocenters were clustered, practically repeating



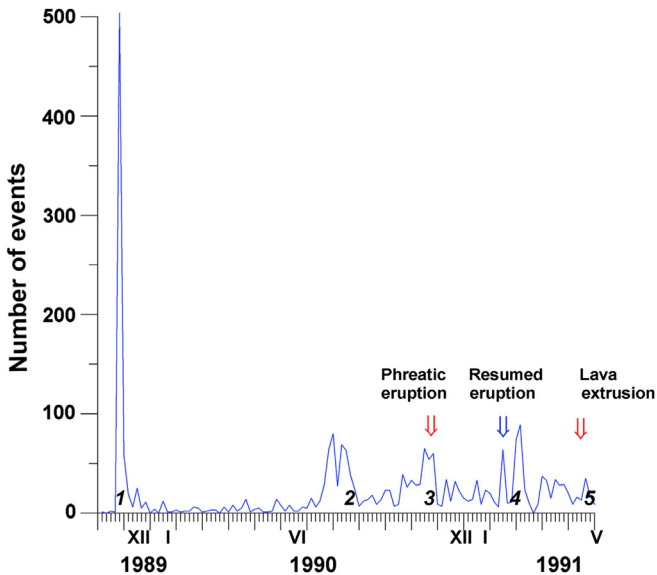


**FIGURE 7.8** Temporal variations of located volcano-tectonic earthquakes (number of events per 5 days) before and during the 1990–91 eruption sequence at Unzen volcano, Japan. Arrows show the main eruptive events. Numbers correspond to the swarm of earthquakes. Data from the catalog of Shimabara Earthquake and Volcano Observatory.

the locations from swarm to swarm, within the long inclined zone that began at the depth of 15 km at a distance of about 20 km from the volcano and terminated at the surface about 2 km from the summit of the volcano (Fig. 7.9C and D).

Swarm 3 (Fig. 7.9B) preceded the phreatic eruption of November 17, 1990. It began on November 13 with a sharp increase in the number of events and continued for about 25 hours (Fig. 7.10A). The phreatic eruption occurred after the cessation of seismic activity. The resumed eruption, which began on February 12, 1991, and was represented by a sequence of phreatic and phreatomagmatic explosions, was not preceded by volcano-tectonic seismicity. Swarm 4 began after the phreatic explosion and was accompanied by a sequence of phreatomagmatic explosions.

Swarm 5 occurred just before the lava extrusion of May 20, 1991. It began on May 18 and reached its peak in the number of events on May 19, 1991. The lava extrusion was observed just after the ceasing of the swarm (Fig. 7.10B). This swarm was the smallest of the five seismic sequences and it was practically at the background level of seismicity (Fig. 7.8). It was characterized by a small number of shallow seismic events with the majority of earthquake foci located at depths from 6 to 15 km (Fig. 7.9D).

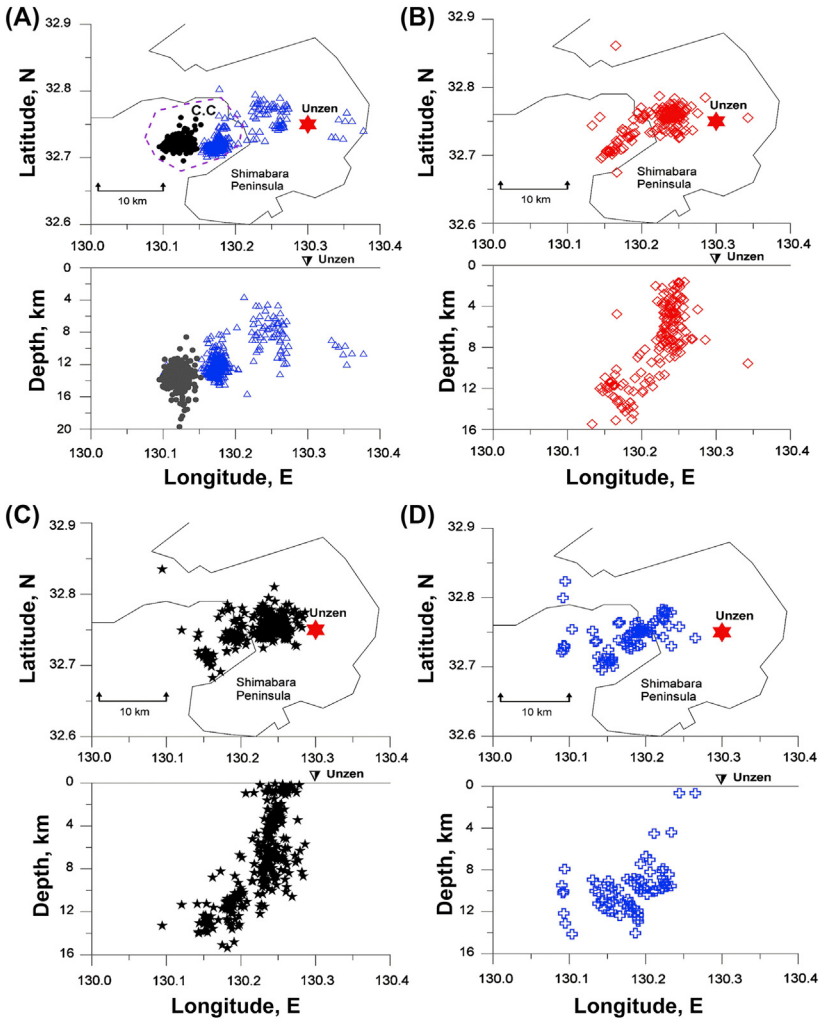


**FIGURE 7.9** Epicenters and cross-sections for the swarms of volcano-tectonic earthquakes that occurred during 1989–91 at Unzen volcano, Japan. The swarms that were not followed by volcanic activity are shown in Fig. A (circles, swarm 1; triangles, swarm 2). The swarms associated with volcanic eruptions are shown in Fig. B (swarm 3), C (swarm 4), and D (swarm 5). Chijima caldera (C.C.) is shown by dashed line. Data from the catalog of Shimabara Earthquake and Volcano Observatory.

The characteristic feature of the Unzen seismicity was the absence of volcano-tectonic events just beneath the volcano crater. Umakoshi et al. (2001) suggested that it was a zone of high-temperature, ductile bodies, or shallow magma chamber.

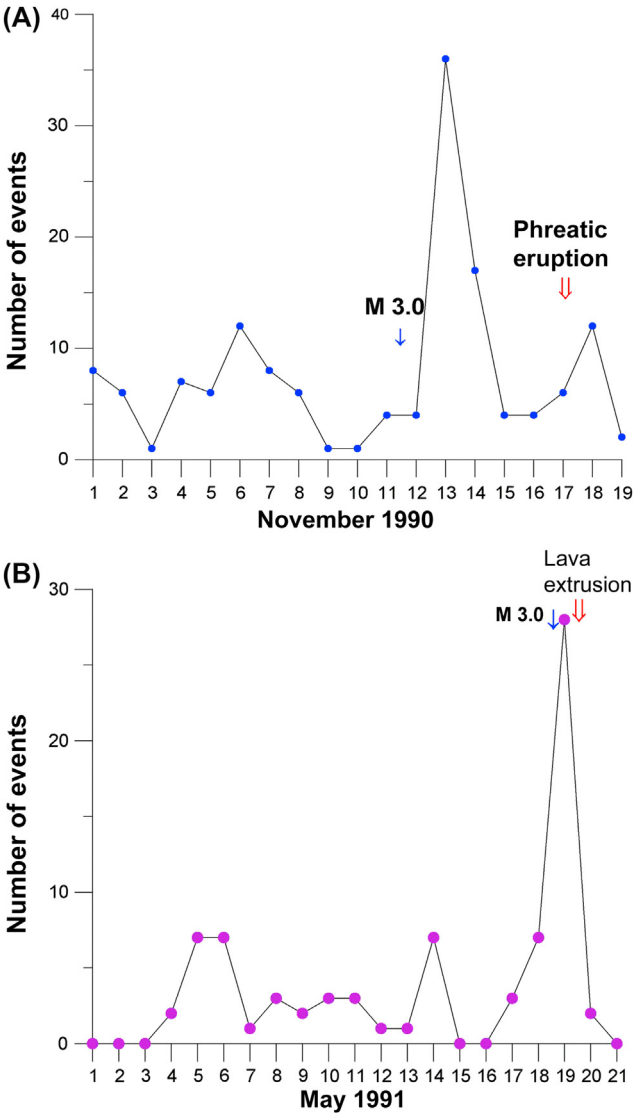


**FIGURE 7.10** Variations of the daily number of earthquakes before the 17 November 1990 phreatic explosion (A) and the 20 May 1991 lava extrusion (B) at Unzen volcano, Japan. Data from the catalog of Shimabara Earthquake and Volcano Observatory.

After the termination of lava activity in February 1995, volcano-tectonic earthquakes were observed along the submeridional structure crossing the focal zone of swarm 1 at a distance of about 20 km from the volcano (Fig. 7.11). They were located at a depth range from 5 to 17 km.

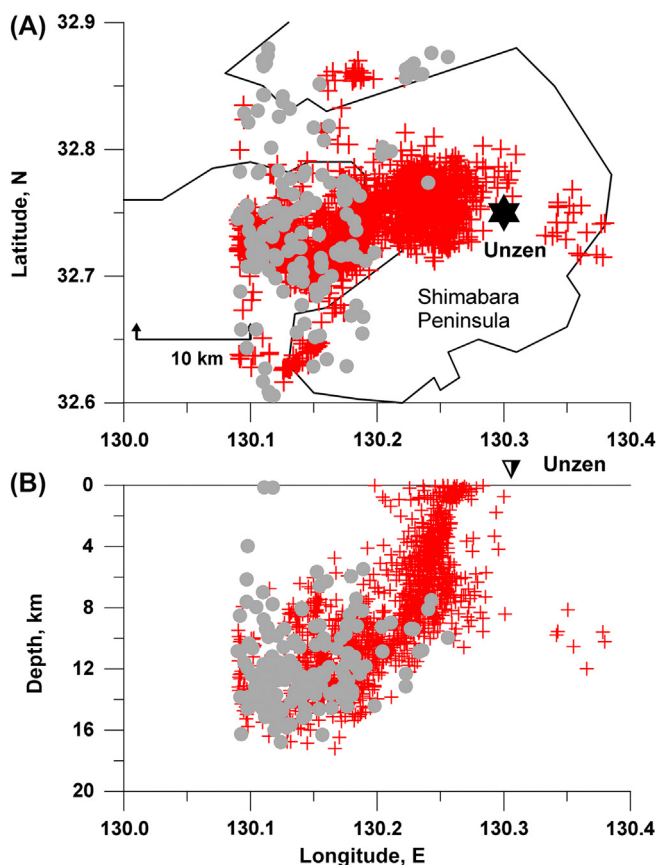


**FIGURE 7.11** The epicenters (A) and hypocenters (B) of the post-eruption volcano-tectonic earthquakes (circles) at Unzen volcano, recorded from February 1995 to the end of 1999. Crosses show the earthquakes recorded during the eruptive activity. Data from the catalog of Shimabara Earthquake and Volcano Observatory.

#### 7.1.4 Pinatubo Volcano, Luzon

Mount Pinatubo is a composite volcano of 1745 m elevation and is situated in the Luzon volcanic arc. Its unrest began after about 500-year calm on April 2, 1991, after about 20 days of locally felt earthquakes related to steam explosions. The unrest culminated 10 weeks later in the great explosion of June 15 that produced about  $10 \text{ km}^3$  of porous, pumiceous deposits and formed a caldera of 2.5 km in diameter. Within the 20th century, it was the second-largest eruption after the 1912 Novarupta-Katmai (Alaska) eruption (Wolfe and Hoblitt, 1996).

Seismic observations began only in May; therefore, the preceding earthquake swarm was not recorded. Nevertheless, the seismic activity, that preceded the June 7 shallow lava extrusion at the summit and following the June 15 great explosion, was located by a seismic network. It allowed for forecasting the June 15 climactic eruption (Harlow et al., 1996).

Figs. 7.12–7.14 show the development of the seismic activity from May 7, 1991, to June 12, 1991. Seismic events were recorded within two clusters: (1) just beneath the summit and (2) 5 km NW of the summit (Fig. 7.12). The second group of seismic foci occurred mainly within the depth range from 7 to 3 km while the first group of events was more shallow, within the range of 4 km below sea level (b.s.l.) to 1 km above sea level (a.s.l.). The events from the distant zone were dominant from May 21 to June 5. Seismicity of zone I beneath the volcano began to increase from May 30, and an intensive swarm of low magnitude (from 0 to 1) was recorded there from June 6 to June 7 (Fig. 7.13). This swarm culminated in the June 7 lava dome extrusion. Just after the beginning of dome growth, the number of earthquakes sharply decreased, and the June 12 explosion occurred in the background of low-level seismicity for both seismic zones.

Fig. 7.14 shows that the seismic foci, that occurred beneath the volcano from May 8 to June 5, became shallower before the June 6–June 7 swarm (and the lava dome extrusion), changing the depths from 3–5 to 0–2 km. The intense seismic activity, which occurred from June 12 to June 15, was practically off the scale. Several of the seismic stations were destroyed, while the remaining were very often saturated (Harlow et al., 1996). The 9-hour climactic sequence of explosions of June 15 was accompanied by large volcano-tectonic earthquakes. Two of them, recorded at 18:41 and 19:15, reached magnitude  $M_w$  5.7 and 5.6, respectively (Harvard CMT).

The location of post-explosion seismic events became possible on 29 June (Mori et al., 1996). The distribution of seismic foci of this period sharply changed compared with the previous one shown in Fig. 7.12. The seismically active zones extended in space and depth (Fig. 7.15). The zone beneath the volcano spread to the west within volcanic edifice and occupied all beneath volcano space filling it up to the depth of 25 km, about 20 km deeper than before the June 15 explosion. The NW seismic zone became also deeper and larger in area.

## 7.2 VOLCANO-TECTONIC EARTHQUAKES ASSOCIATED WITH FLANK ERUPTIONS

We have only three cases; all are related to the activity of the Usu volcano, Hokkaido.

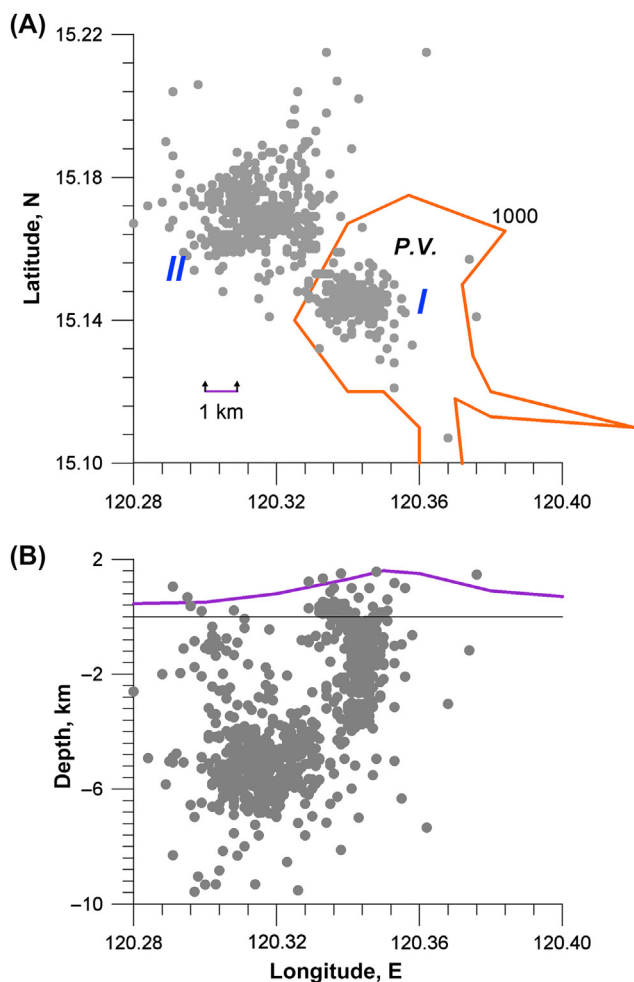


**FIGURE 7.12** Epicenters (A) and hypocenters (B) of earthquakes located from May 7, 1991, to June 12, 1991, near Pinatubo volcano, Philippines. The three-dimensional velocity model was used for location (Harlow et al., 1996). PV denotes Pinatubo volcano. I and II mark two clusters of earthquakes. The 1000-m contour of the volcano's relief and the zero-line of sea-level depth are shown. Data from Mori, J., White, R.A., Harlow, D.H., Okubo, P., Power, J.A., Hoblitt, R.P., Laguerta, E.P., Lanuza, A., Bautista, B.C., 1996. Volcanic earthquakes following the 1991 climactic eruption of Mount Pinatubo: strong seismicity during a waning period. In: Newhall, C. G., Punongbayan, R.S. (Eds.), *Fire and Mud. Eruptions and Lahars of Mount Pinatubo, Philippines*. University of Washington Press, Seattle, pp. 339–350.

## 7.2.1 Usu Volcano, Hokkaido

### 7.2.1.1 Eruption of 1910

The eruption of the cryptodome Meiji-shinzan began at the northern foot of the volcano on July 25 (see), after 4 days of precursory earthquakes

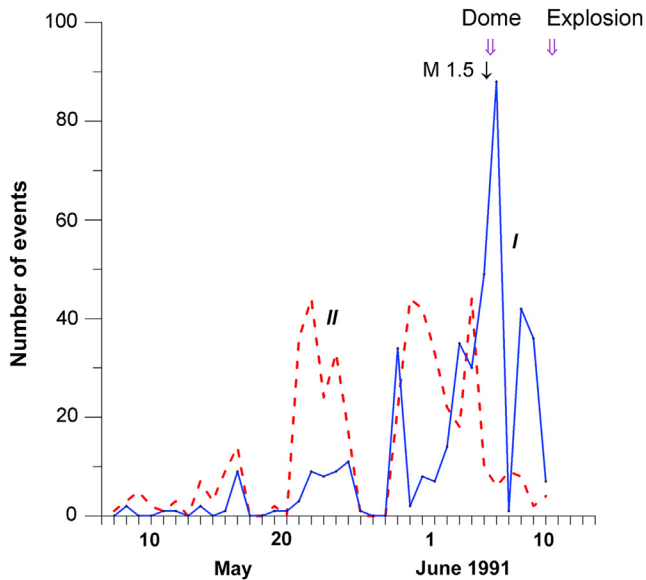


**FIGURE 7.13** Temporal variations in the daily number of earthquakes located from May 7, 1991, to June 12, 1991, beneath the Pinatubo volcano (I) and 5 km to NW from the volcano (II). The double arrows show the eruptive events; the single arrow shows the maximum magnitude earthquake. I and II are two clusters of earthquakes. Data from Mori, J., White, R.A., Harlow, D. H., Okubo, P., Power, J.A., Hoblitt, R.P., Laguerta, E.P., Lanuza, A., Bautista, B.C., 1996. Volcanic earthquakes following the 1991 climactic eruption of Mount Pinatubo: strong seismicity during a waning period. In: Newhall, C.G., Punongbayan, R.S. (Eds.), *Fire and Mud. Eruptions and Lahars of Mount Pinatubo, Philippines*. University of Washington Press, Seattle, pp. 339–350.

(Fig. 7.16). Nearly 40 craterlets were formed along the arcade line approximately 3 km long during 10 days of phreatic explosive activity. The eruption caused an upheaval of approximately 90 m above the original terrace (Yokoyama and Seino, 2000).

The felt shakes began on July 21. On July 22, about 25 shocks were felt at a distance of 8 km from the volcano. Fig. 7.16B shows the 6-hour frequency of earthquakes recorded by a horizontal pendulum seismograph (magnification  $\times 30$ ) that was installed at the meteorological observatory of Sapporo, at a distance of 70 km from the volcano. The number of earthquakes increased rapidly and reached its maximum 13 hours before the beginning of the eruption. The strongest earthquake ( $M_w$  5.4; Zobin, 2001) was recorded 30 hours before the beginning of the eruption. The eruption began on the background of a gradual decrease in the number of earthquakes (Omori, 1911–1913). Fig. 7.16 shows also the number of earthquakes that were felt at Nishi-Monbets village, at a distance of 8.4 km from the volcano (Fig. 7.16A). The remarkable similarity between the curves is seen.

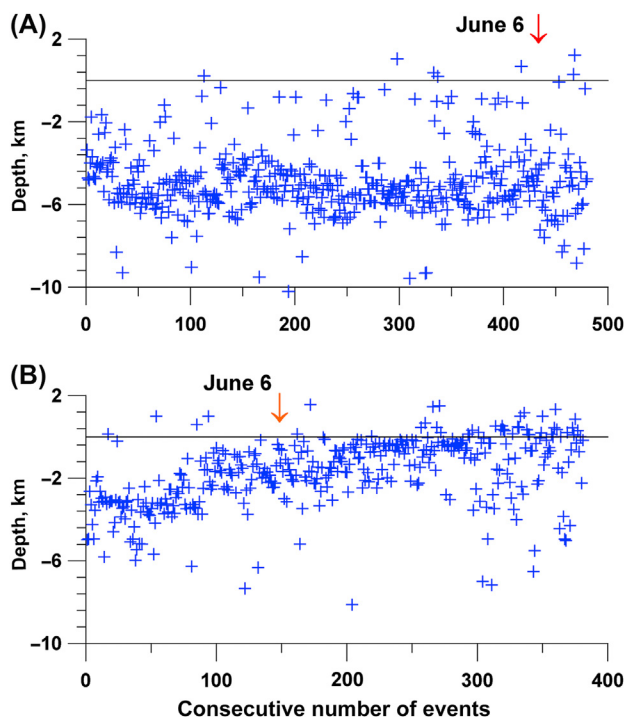


**FIGURE 7.14** Variations in the focal depths of the events from cluster II situated 5 km to NW from the volcano (A) and from cluster I beneath the volcano Pinatubo (B) from May 7, 1991, to June 12, 1991. Zero-line of sea-level depth is shown. The beginning of the June 6 to June 7 June swarm is shown by the arrow. Data from Mori, J., White, R.A., Harlow, D.H., Okubo, P., Power, J.A., Hoblitt, R.P., Lagueria, E.P., Lanuza, A., Bautista, B.C., 1996. Volcanic earthquakes following the 1991 climactic eruption of Mount Pinatubo: strong seismicity during a waning period. In: Newhall, C.G., Punongbayan, R.S. (Eds.), *Fire and Mud. Eruptions and Lahars of Mount Pinatubo, Philippines*. University of Washington Press, Seattle, pp. 339–350.

### 7.2.1.2 Eruption of 1944

It was the birth of a new volcano Showa-Shinzan on the flank of Usu (see). The seismic activity in the area of the forthcoming volcano began on December 28, 1943, with numerous earthquakes. Fig. 7.17 shows the variations in the preeruption sequence of earthquakes that continued for 6 months. This figure is based on the catalog of earthquakes felt in the area of the Usu volcano that was prepared by Mimatsu (1995).

The maximum number of events (up to 120 daily shocks) was recorded during the first 3 days of the activity. The maximum magnitude event ( $M$  5; Yokoyama and Seino, 2000) occurred on January 9, 1944, in 12 days



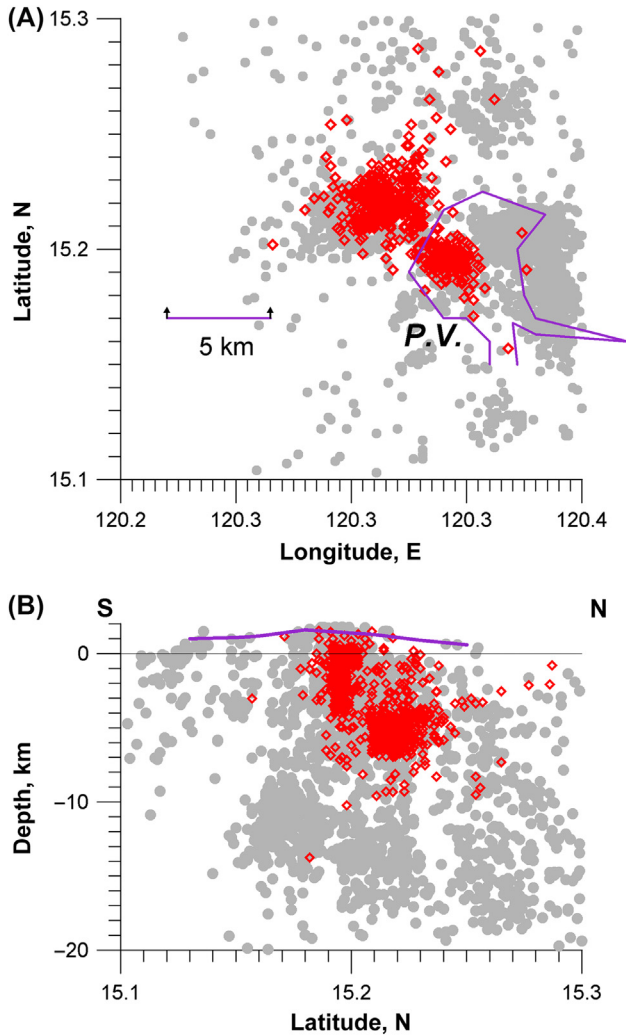


**FIGURE 7.15** Following June 15, 1991 seismic activity in the region of Pinatubo observed from June 26, 1991, to August 18, 1991. Epicenters (A) and hypocenters (B) of earthquakes (N–S cross-section). The earthquakes, recorded before the June 15 explosions, are shown by diamonds; the earthquakes, recorded after the June 15 explosions, are shown by circles. The 1000-m contour of Pinatubo volcano (PV) is shown. Data from Mori, J., White, R.A., Harlow, D.H., Okubo, P., Power, J.A., Hoblitt, R.P., Laguerta, E.P., Lanuza, A., Bautista, B.C., 1996. Volcanic earthquakes following the 1991 climactic eruption of Mount Pinatubo: strong seismicity during a waning period. In: Newhall, C.G., Punongbayan, R.S. (Eds.), *Fire and Mud. Eruptions and Lahars of Mount Pinatubo, Philippines*. University of Washington Press, Seattle, pp. 339–350.

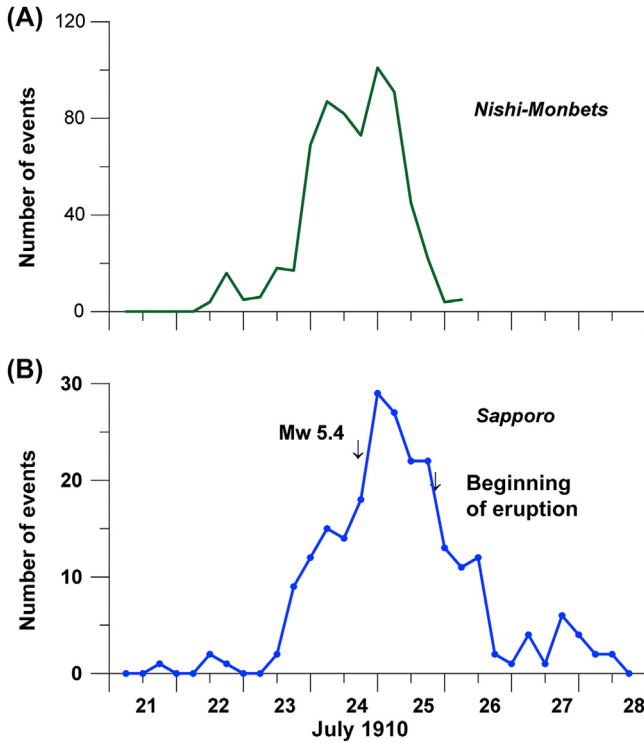


**FIGURE 7.16** Variations in the 6-h number of earthquakes preceding the 1910 flank eruption of Usu volcano. (A) From macroseismic data at a distance of 8.5 km from the new volcanic site; (B) Recorded by a seismic station at a distance of 70 km from the new volcanic site. *Data from Omori, F., 1911–1913. The Usu-san eruption and earthquake and elevation phenomena. Bulletin of the Imperial Earthquake Investigation Committee 5, 1–38.*

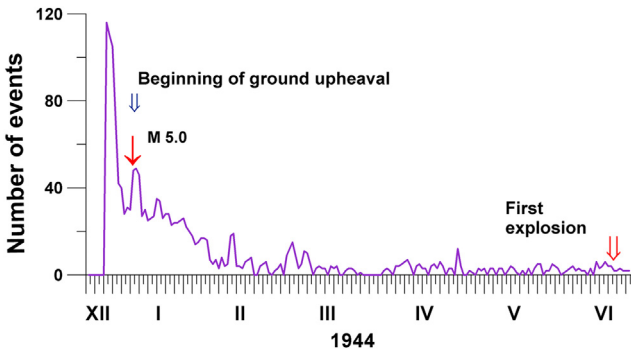


**FIGURE 7.17** Variations in the daily number of felt earthquakes preceding the June 23, 1944 flank eruption at Usu volcano. The *double arrows* show the beginning of the ground upheaval in the area of forthcoming eruption and the initial stage of the eruption; the *single arrow* shows the appearance of the maximum magnitude earthquake of this sequence. *Data from Mimatsu, M., 1995. Showa-Shinzan Diary. Suda Seihan, Sapporo. 179 p.*

after the beginning of the activity, and was located by a three-station network at the depth of 11 km (Kizawa, 1958). The appearance of this earthquake coincided with the beginning of the upheaval of the overlying ground. Eventually, the upheaval area started to spread to the north, reaching up to a height of 50 m in some areas. Seismic activity that accompanied this deformation process gradually decreased from January 1944 to February 1944. It stayed then at the background level of 0–10 daily shocks in March–June.

Many fissures started to appear in the surrounding farmlands. The longest fissure recorded was about 600 m long. On June 23, 1944, about 6 months after the beginning of the seismic activity, smoke started to come out of the fissures, and shortly after a huge eruption followed accompanied by a thunderous roar. By the end of October 1944, a total of seven explosion craters were formed. The new lava dome of Showa-Shinzan continued to grow up to September 1945 (Mimatsu, 1995).

### 7.2.1.3 Eruption of 2000

The 2000 eruption began on the northwestern flank of the volcano (Fig. 7.18) on March 31 at 13:07 (local time), following about 4 days of intensive seismic activity and a high-rate deformation of the volcano surroundings (Ichiyangi, 2002; Takahashi et al., 2002). More than 60 explosion craters were formed in March and April by the sequence of phreatomagmatic and phreatic explosions within two new eruptive centers: Nishi-yama (N in Fig. 7.18) and Kompira-yama (K in Fig. 7.18). The initial phreatomagmatic eruption on March 31 was the largest and it produced a tephra fall with a small pyroclastic flow. The volume of dacitic ash and pumice was about  $0.1 \text{ km}^3$ . The eruption was accompanied by an upheaval process that reached about 75–80 m at the Nishi-yama cryptodome (Jousset et al., 2003). After the summer of 2000, the intensity of upheaval decayed gradually.

About 1200 earthquakes were located by JMA from March 28 to April 1; 339 of them with magnitudes ranging from 2.5 to 4.6 (Ueno et al., 2002). The regional seismic network of JMA and the Institute of Seismology and Volcanology of Hokkaido University comprised 16 short-period stations, situated at a distance of 1–100 km from the volcano (Ueno et al., 2002). The network allowed locating the earthquakes with mean errors of  $\pm 0.6 \text{ km}$  for latitude and  $\pm 0.7 \text{ km}$  for longitude. The mean depth errors were equal to  $\pm 2.2 \text{ km}$ . The earthquake magnitudes  $M_{\text{JMA}}$  (or  $M$ ) were estimated for 948 events.

Seismic activity, preceding the 2000 Usu eruption, began on March 27 at 20 hours (local time) with unlocated low-magnitude earthquakes ( $M < 1$ ). The first located earthquake of magnitude  $M$  1.0 occurred on March 28 at 00:10 at a depth of 5.7 km. During the first 40 hours of activity, the frequency of events was rather low, not exceeding 10 located events per hour

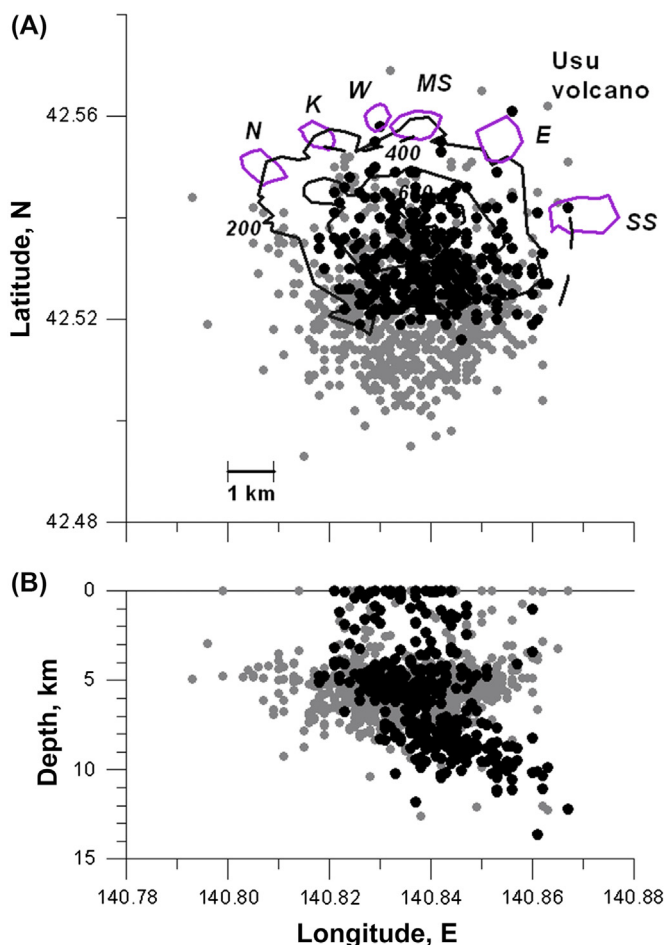


**FIGURE 7.18** Epicenters (A) and hypocenters (B) of earthquakes that occurred at Usu volcano from March 28, 2000, to April 1, 2000. The events of March 28 to March 29 are shown by *dark symbols*; the events of March 30 to March 31 are shown by *light symbols*. The 200, 400, and 600 m contours of the Usu volcano are shown by *lines*. The centers of flank eruptions of the Usu volcano are shown by the ovals with letters. *N* and *K* are the 2000 centers; *MS* and *SS* are the centers of eruptions during 1910 and 1944, respectively; the centers *W* and *E* were not dated. The *broken lines* show the lateral extent of volcanic activity during the 1910 and 1943–44 flank eruptions. The positions of flank eruptive centers and the lines of lateral extent of volcanic activity are shown according to Yokoyama and Seino (2000). Data from the catalog of Japan Meteorological Agency.

(Fig. 7.19). The maximum magnitudes of earthquakes were 3.2–3.4. The number of earthquakes and their magnitudes increased sharply at the end of March 29 and reached their maximum between 9 and 12 hours of March 30. Then seismic activity gradually decreased and practically ceased on April 2. The three largest earthquakes of the swarm occurred during the maximum

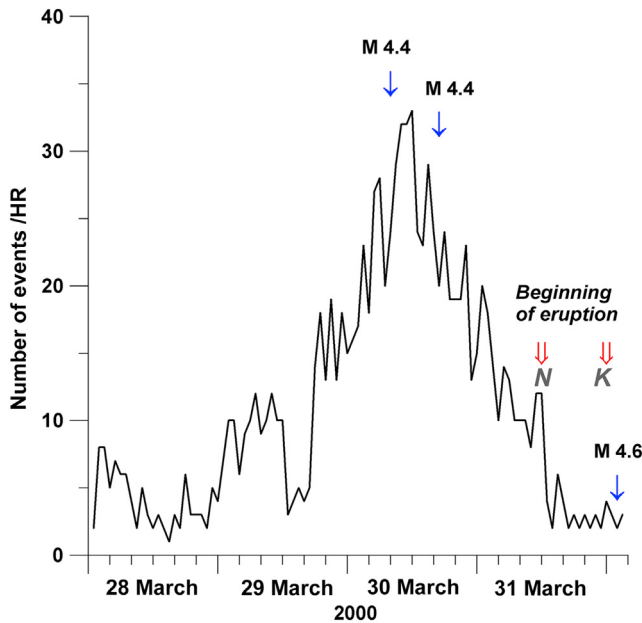


**FIGURE 7.19** Temporal developments of the 2000 earthquake swarm at Usu (JMA catalog). The *single arrows* show the position of the largest earthquakes; the *double arrows* show the beginning of the eruption in the Nishi (N) and Kompira (K) centers of the eruption. Data from the catalog of Japan Meteorological Agency.

seismic energy release ( $M$  4.4, March 30, 09:12 and 17:13) and at the final stage of the swarm, 14 hours after the beginning of the eruption ( $M$  4.6;  $M_w$  4.9, April 1, 03:12; [Zobin et al., 2005](#)).

The earthquake epicenters occupied an area of about  $30 \text{ km}^2$ , covering the summit zone and its southern and southwestern flanks of the volcano ([Fig. 7.18](#)). Initially, the earthquakes occurred within the zone situated beneath the summit and southern slope of the volcano. Then, the seismic foci clustered along two linear zones intersecting to the south of the initial epicentral zone. The foci were recorded within the depth range between 0 and 13 km, but clustered mainly from 4 to 7 km. The earthquake epicenters stopped at a distance of about 1–3 km from the sites of the forthcoming eruption.

The alert of the forthcoming eruption was issued by JMA on the morning of March 29; by that night, more than 15,000 people from the towns around the volcano were evacuated (JMA, 2003).

The March 31 phreatomagmatic explosion from the *N* vents occurred on the northwestern flank of the volcano about 4 km from the summit of Usu, at the end of the seismic swarm ([Fig. 7.19](#)) and just beyond the zone of the earthquake epicenters ([Fig. 7.18](#)).

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## RELEVANT WEBSITES

Harvard Centroid Moment Tensors. <http://www.globalcmt.org/CMTsearch.html>.

Japan Meteorological Agency (JMA, 2003). <https://www.jma.go.jp/jma/indexe.html>.

## Chapter 8

# General Properties of Volcano-Tectonic Earthquake Swarms

This chapter discusses the general properties of volcano-tectonic earthquake swarms. First, the properties of the earthquake swarm inferred from the data presented in Chapters 5–7 are generalized. Then, these data would be supplemented and compared with the published data in this field.

### 8.1 COMPARATIVE CHARACTERISTICS OF VOLCANO-TECTONIC EARTHQUAKE SWARMS OCCURRING AT BASALTIC, ANDESITIC AND DACITIC VOLCANOES

We passed through a sequence of seismovolcanic episodes discussed in Chapters 5–7. [Table 8.1](#) presents the characteristics of seismic sequences of volcano-tectonic earthquakes observed in the 20th and the beginning of the 21st centuries at 21 volcanoes of the world before their 41 eruptions. These data are not representative of the volcanic activity occurring on the Earth during the last 100 years; it is only a small sample showing the most studied and/or most remarkable examples of seismic activity at volcanoes. Nevertheless, the general properties (or tendencies) of seismic swarms, associated with preliminary volcanic activity as well as the descriptions of the posteruption seismicity presented in Chapters 5–7, can be discussed. To do these, the following characteristics are selected: temporal variations in seismic activity, spatial distribution of seismic foci, duration of seismic activity before volcanic event, position of volcanic event in seismic sequence, posteruption seismicity, and dependence of these characteristics on the type of magma feeding the eruption. The last section is devoted to some regularity in the seismic swarm activity associated with the reawakening of andesitic and dacitic volcanoes.

#### 8.1.1 Temporal Variations

The temporal variations in seismic activity were of three types: (I) one-peak sequences with the maximum peak at the beginning of the sequence, (II) multi-peak sequences with the maximum peak at the beginning of the sequence, or (III) one- and multi-peak sequences with the maximum peak at the end of the sequence ([Fig. 8.1A](#)). The first type of sequence is rare ([Fig. 8.1B](#)), and it was observed



**TABLE 8.1** Data on Volcano-Tectonic Earthquake Sequences Recorded Before Eruptions.

| Volcano,<br>Region              | Type of<br>Magma | Year | Type of<br>Eruption | VEI | Duration of<br>Preliminary<br>Earthquake<br>Sequence (h) | Depth<br>(km) | Range of<br>Magnitudes | Type of<br>Earthquake<br>Sequence | Position of<br>Eruption<br>Event |
|---------------------------------|------------------|------|---------------------|-----|----------------------------------------------------------|---------------|------------------------|-----------------------------------|----------------------------------|
| Kilauea<br>Hawaii               | Basaltic         | 1969 | 4                   | 0   | 5                                                        | 2–10          | 0–3.5                  | I                                 | IV                               |
|                                 |                  | 1974 | 1.4                 | 0   | 9                                                        | 0–4           | 0–3.4                  | III                               | IV                               |
|                                 |                  | 1982 | 1                   | 0   | 2.5                                                      | 1–3           | 0–3.2                  | I                                 | III                              |
| El Hierro,<br>Canary Island     | Basaltic         | 2011 | 5                   | 2   | 768                                                      | 10–20         | 0–4.3                  | II                                | III                              |
| Etna, Sicily                    | Basaltic         | 1983 | 2                   | 1   | 28                                                       | (–)3–15       |                        | III                               | IV                               |
|                                 |                  | 1984 | 1                   | 2   | 0                                                        |               |                        | —                                 | —                                |
|                                 |                  | 1991 | 2                   | 2   | 14                                                       | 0–10          | 0–4.4                  | II                                | I                                |
|                                 |                  | 2001 | 2                   | 2   | 120                                                      | (–)2–5        | 0–3.9                  | I                                 | III                              |
| Oshima, Izu<br>Island           | Basaltic         | 1986 | 1                   | 3   | 0                                                        |               |                        |                                   |                                  |
|                                 |                  | 1986 | 2                   | 3   | 2                                                        | 0–5           | 0–5.9                  | —                                 | III                              |
| Klyuchevs-<br>koy,<br>Kamchatka | Basaltic         | 1951 | 2                   | 2   | 96                                                       |               | 0–3.1                  | II                                | IV                               |
|                                 |                  | 1966 | 2                   | 3   | 84                                                       |               | 0–3.7                  | II                                | IV                               |

|                              |           |      |       |   |         |      |       |     |     |
|------------------------------|-----------|------|-------|---|---------|------|-------|-----|-----|
| New Tolbachik, Kamchatka     | Basaltic  | 1975 | 4     | 4 | 216     | 0–16 | 0–4.9 | II  | IV  |
| Dabbahu, Ethiopia            | Basaltic  | 2007 | 4     | 3 | 340–600 | 2–12 | 0–5.4 | II  | IV  |
| Fernandina, Galapagos Island | Basaltic  | 1968 | 1     | 4 |         |      | 0–5.2 | —   | —   |
| Teishi Knoll, Izu Island     | Basaltic  | 1989 | 5     |   | 216     | 0–12 | 0–5.5 | II  | IV  |
| Miyakejima, Izu Island       | Basaltic  | 2000 | 2, 5  |   | 70      | 1–4  |       | I   | II  |
|                              |           | 2000 | 1a, 5 | 3 | 1200    | 0–20 | 1–6.4 | I   |     |
| Bezymianny, Kamchatka        | Andesitic | 1955 | 1     | 5 | 408     |      |       | III | II  |
|                              |           | 1956 | 1     | 5 | 4320    |      | 0–5.2 | II  | III |
|                              |           | 1959 | 1     | 1 | 48      |      | 0–1.3 | I   | II  |
|                              |           | 1960 | 1     | 2 | 70      |      | 0–1.7 | I   | II  |
| Sheveluch, Kamchatka         | Andesitic | 1964 | 1     | 4 | 216     |      | 0–5.2 | III | II  |
|                              |           | 1993 | 1     | 3 | 500     |      |       | III | II  |

(Continued)

**TABLE 8.1 (Continued)**

| Volcano,<br>Region                | Type of<br>Magma | Year | Type of<br>Eruption | VEI | Duration of<br>Preliminary<br>Earthquake<br>Sequence (h) | Depth<br>(km) | Range of<br>Magnitudes | Type of<br>Earthquake<br>Sequence | Position of<br>Eruption<br>Event |
|-----------------------------------|------------------|------|---------------------|-----|----------------------------------------------------------|---------------|------------------------|-----------------------------------|----------------------------------|
| El Chichon,<br>Mexico             | Andesitic        | 1982 | 1                   | 5   | 620                                                      | 0–20          | 0–4.0                  | II                                | III                              |
| Volcán de<br>Colima,<br>Mexico    | Andesitic        | 1998 | 1                   | 3   | 600                                                      | (–)4–0        | 0–2.7                  | III                               | III                              |
| Soufrière<br>Hills,<br>Montserrat | Andesitic        | 1995 | 1                   | 3   | 30                                                       | (–)0.5–4      | 0–1.4                  | III                               | III                              |
|                                   |                  | 1996 | 1                   |     | 600                                                      | (–)0.5–3      | 0–1.9                  | III                               | III                              |
| Popocatép-<br>etl, Mexico         | Andesitic        | 1994 | 1                   | 2   | 8                                                        | (–5)–2        | 1–3.0                  | II                                | II                               |
| Sakurajima,<br>Kyushu             | Andesitic        | 1914 | 2                   | 4   | 80                                                       |               | 0–7.0                  | III                               | IV                               |
| Mount St.<br>Helens,<br>Cascades  | Dacitic          | 1980 | 1                   | 5   | 80                                                       | 0–20          | 0–4.4                  | II                                | II                               |
|                                   |                  | 1980 | 1                   | 5   | 1300                                                     | 0–10          | 0–5.7                  | II                                | III                              |
|                                   |                  | 2004 | 1                   | 2   | 570                                                      | 0–2           | 1–3.8                  | II                                | III                              |

|                          |         |      |   |   |      |          |       |     |     |
|--------------------------|---------|------|---|---|------|----------|-------|-----|-----|
| Usu,<br>Hokkaido         | Dacitic | 1910 | 2 | 2 | 108  |          | 0–5.4 | III | III |
|                          |         | 1944 | 2 | 2 | 700  |          | 0–5.0 | II  | III |
|                          |         | 1944 | 2 | 2 | 4300 |          |       | II  | IV  |
|                          |         | 1977 | 1 | 3 | 32   | 0–2      | 0–3.7 | III | II  |
|                          |         | 2000 | 2 | 2 | 90   | 0–15     | 1–4.6 | I   | III |
| Unzen,<br>Kyushu         | Dacitic | 1990 | 1 | 1 | 25   | 1–15     | 0–3.0 | I   | IV  |
|                          |         | 1991 | 1 | 1 | 48   | 0–15     | 0–3.0 | I   | III |
| Pinatubo,<br>Philippines | Dacitic | 1991 | 1 | 6 | 160  | (–)1–1.6 | 0–1.6 | 3   | II  |

*Notes:* 1, central eruption; 1a, caldera collapse; 2, flank eruption; 3, radial fissure; 4, regional fissure; 5, submarine eruption. For the codes of earthquake sequence (I–III) and of the position of volcanic events (I–IV; see in text). Symbol (–) before the depth means the depth above sea level.

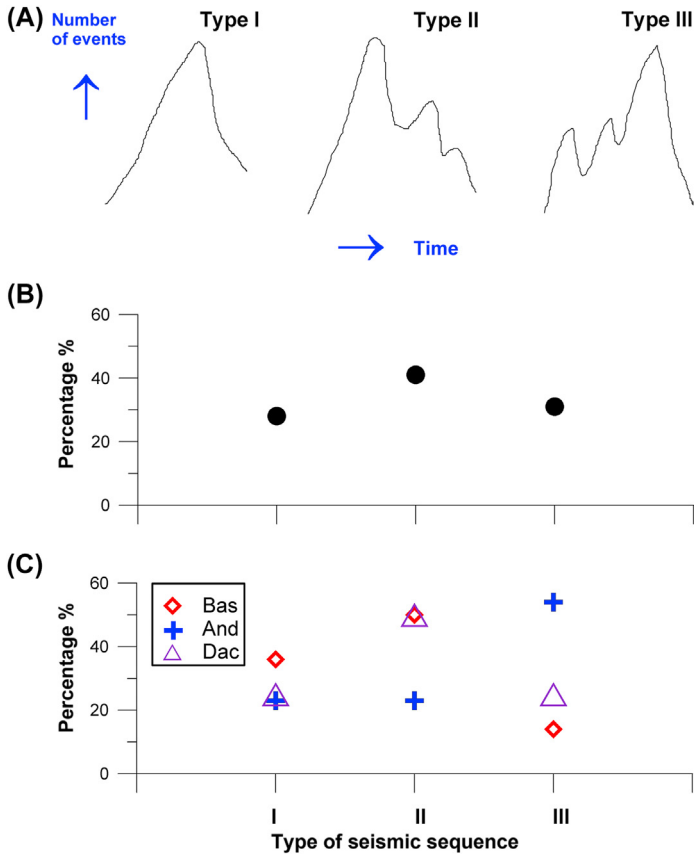


**FIGURE 8.1** Three typical curves of swarms development (A) and the distribution of three corresponding types of volcano-tectonic earthquake swarms for all events of Table 8.1 (B) and for basaltic, andesitic, and dacitic volcanoes (C). The three types of volcano-tectonic earthquake swarms are (I) one-peak sequences with the maximum peak at the beginning of the sequence, (II) multi-peak sequences with the maximum peak at the beginning of the sequence and (III) one- and multi-peak sequences with maximum peak at the end of the sequence.

only for 10 from a total of 37 cases. The second type of sequence, more characteristic, was observed for 16 cases, and the third was observed for 12 cases.

For basaltic and dacitic volcanoes, the most popular is a seismic sequence with the maximum peak at the beginning of the sequence (Type II), while for andesitic volcanoes, the maximum number of seismic events for the majority of seismic sequences is observed at the final stage of the swarm (Type III; Fig. 8.1C).

### 8.1.2 Spatial Distributions

Generally, the seismic events cluster within or below the volcanic edifice within a depth range from 20 km to the summit of the volcano. The seismic

events of preliminary swarms occupy a greater area than the site of the forthcoming eruption. At the initial stage of seismic activity, the events occurring at the depths of 10–20 km were frequently observed together with the shallower events. Before an eruption, the earthquakes cluster at shallower depths. It was observed for New Tolbachik volcanoes (Figure 5.17 from Chapter 5), Mount St. Helens (Figure 7.2 from Chapter 7), Pinatubo (Figure 7.10), and Volcán de Colima (Figure 6.20 from Chapter 6), among others. Very often the earthquakes originate not only beneath the volcano but also along a system of local faults adjusting the volcano. Examples of well-located earthquakes show that during the initial stage of seismic activity before an eruption, the seismic events were observed along the local faults, such as St. George's Hill of Soufrière Hills volcano (Figure 6.14 from Chapter 6), the fault 5 km NW of Pinatubo volcano (Figure 7.12); and the fault SE of Mount St. Helens (Figure 7.2 from Chapter 7). The adjacent fault events had a greater depth than the events originating beneath the volcano. Just before an eruption, seismic activity began to concentrate at shallower depths beneath the volcanic edifice.

The important feature of the spatial distribution of volcano-tectonic earthquakes is the migration of seismic foci (or seismic clusters) for a distance of 10–20 km. The focal migration was observed before the forming of the eighth cone of the New Tolbachik volcanoes (Figure 5.21 from Chapter 5). Earthquakes migrated for 6 days along the fault at a distance of about 15 km from the site of the formation of the first seven cones to the site of the eighth cone.

Multipeak temporal variations in the number of seismic events may be considered as a result of the migration of seismic clusters. Before the submarine eruption of Teishi Knoll, the two peaks in temporal variations of the number of earthquakes corresponded to the clustering of seismic events within two separate zones (Figure 5.28 from Chapter 5 and Figure 5.30 from Chapter 5). One of these zones was active from July 4, 1989, to July 7, 1989, and produced the first peak of frequency of events; the second cluster developed in a longer and narrower zone and produced the second peak of frequency. The same was observed at New Tolbachik volcanoes (Figure 5.19 from Chapter 5 and Figure 5.20 from Chapter 5). During the period from June 27, 1975, to June 30, 1975, the earthquake source zone was larger and deeper and produced the first peak of earthquake frequency; during the period from July 1, 1975, to July 6, 1975, the earthquakes occurred within a longer and narrower zone and produced the second peak of earthquake frequency.

### **8.1.3 Posteruption Seismic Activity**

There is generally no volcano-tectonic seismicity at the beginning of the eruption. Nevertheless, the posteruption seismicity was observed after the eruptions of New Tolbachik volcanoes, Mount St. Helens, Oshima, Unzen, Pinatubo, and El Hierro. The spatial distribution of the posteruption volcano-tectonic earthquakes overfills the areas that were active before and during the eruptive events. For the

New Tolbachik volcanoes (Figure 5.23 from Chapter 5), the epicenters of posteruption earthquakes, that were observed 2 years after the eruption, outline the 100-km-length fault that includes the 12-km new eight cones fissure. They occurred at the same depths as the events of the preliminary earthquake swarm.

For Mount St. Helens, the activated posteruption fault, whose central part was occupied by the volcano, was outlined to the north and to the south of the volcano. The free space between two seismic clusters that occurred separately at the depths from 0 to 10 km and from 16 to 22 km before and during the eruption was filled by the postvolcanic seismic events during 1 year of activity (Figure 7.3 from Chapter 7).

For Unzen volcano, the posteruption events traced a 30-km-long structure that intersects at a depth of 20 km the seismic volume where the first preliminary swarm occurred 1 year before the eruption (Figure 7.11).

In the case of Pinatubo volcano, the posteruption seismic events significantly expanded the seismic volume of preliminary events. The epicentral area and the depths of earthquakes were both increased by a factor of two (Figure 7.15).

The volcano-tectonic earthquakes, that followed the eruption of Oshima volcano, also significantly expanded the seismic volume that was active before the eruption (Figure 5.13 from Chapter 5).

The swarm III of the El Hierro seismic sequence occurred during the eruption period but out of the eruption area, to the north of the swarm I (Figure 5.35 from Chapter 5). It stands out almost as an independent earthquake sequence. The appearance of this swarm beneath the El Golfo depression, 20–30 km to the north of the area of the eruption, with an increase of focal depths of earthquakes and their magnitudes up to *ML* 4.6, may be the result of the subsidence of the crustal structures covering the initial magma chamber.

The spatial characteristics of the posteruption seismicity suggest its relationship with the collapses of magma chambers after the magma release during the eruption.

#### 8.1.4 Duration of Seismic Swarms Before an Eruption

The duration of seismic swarms of the majority (90%) of seismic sequences of Table 8.1 ranges between 10 and 1000 hours (a geometrical mean of  $117 \pm 7$  hours; Fig. 8.2A). These values mainly reflect the properties of andesitic and dacitic volcanoes (Fig. 8.2B); for the majority of basaltic volcanoes, the durations of seismic activity before eruption are shorter (from 1 to 100 hours; a geometrical mean of  $55 \pm 7$  hours).

#### 8.1.5 Position of a Volcanic Event According to the Stage of Volcano-Tectonic Earthquake Swarm

Volcanic events may occur during the different stages of the seismic sequence (Fig. 8.3A): (1) Stage I, at the initial stage of the seismic sequence, (2) Stage II, at the maximum frequency of seismic events, (3) Stage III, at

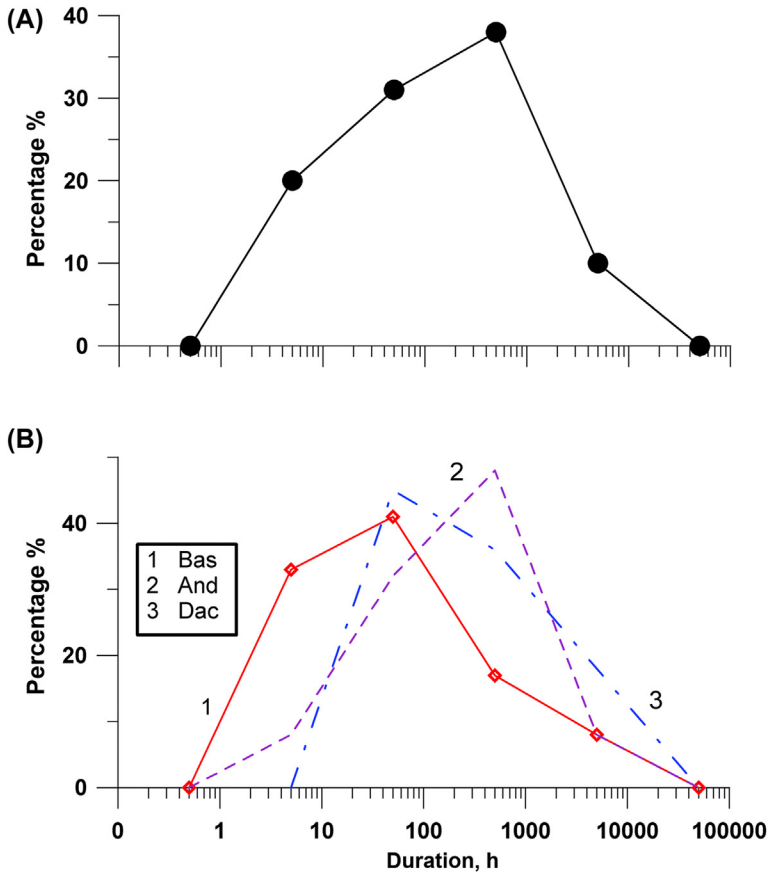


**FIGURE 8.2** Distribution of duration of volcano-tectonic earthquake swarms for all events of Table 8.1 (A) and for basaltic (1), andesitic (2), and dacitic, (3) volcanoes (B).

the ceasing of seismic events, and (4) Stage IV, after ceasing of seismic sequence. The majority of the discussed eruptions occurred while the earthquakes were ceasing (Stage III; Fig. 8.3B). For basaltic volcanoes, the majority of eruptions occur after the end of the seismic sequence, while for andesitic volcanoes, the majority of eruptions occur during the maximum number of volcano-tectonic earthquakes (Fig. 8.3C).

## 8.2 ADDITIONAL DATA ABOUT VOLCANO-TECTONIC EARTHQUAKE SWARM PROPERTIES

### 8.2.1 Size of Volcano-Tectonic Earthquake Swarm Area

The epicentral area of an earthquake swarm is defined as the area of volcano-tectonic earthquake epicenters (magnitude 1–2 and greater) during eruptive



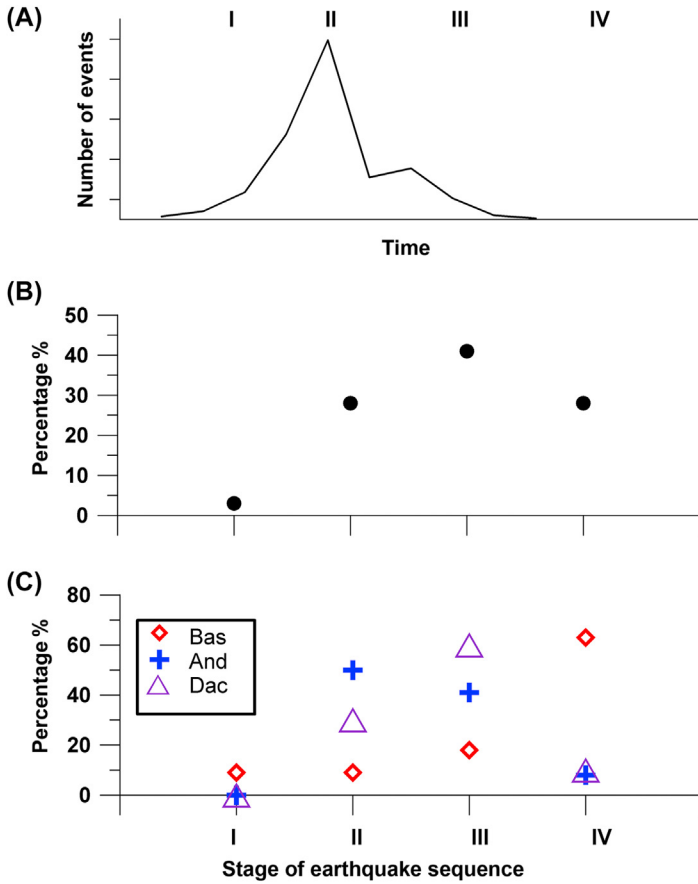


**FIGURE 8.3** Position of volcanic event according to the different stages in volcano-tectonic earthquake swarms for all events of Table 8.1 (B) and for basaltic, andesitic, and dacitic volcanoes (C). The stages are illustrated in (A): the initial stage of seismic sequence (Stage I), stage at the maximum frequency of seismic events (Stage II), at the ceasing of seismic events (Stage III), and after ceasing of seismic sequence (Stage IV).

activity. Areas of epicentral zones vary for different volcanoes from  $25 \text{ km}^2$  for Mount St. Helens to  $9450 \text{ km}^2$  for northern Tonga (Zobin, 1996). Zobin (1996) showed that the areas are larger for volcanoes on small islands and smaller for volcanoes on large islands or on continents. Assuming that the crust is thinner for oceanic volcanoes than for continental volcanoes, it was found that a significant negative correlation exists between the epicentral areas of volcanic earthquakes  $S$  and the crustal thickness  $H$  under the volcano (Fig. 8.4):

$$\log S (\text{km}^2) = 3.6 \times 10^5 H^{-3.1} \quad (\text{km}) \quad (8.1)$$

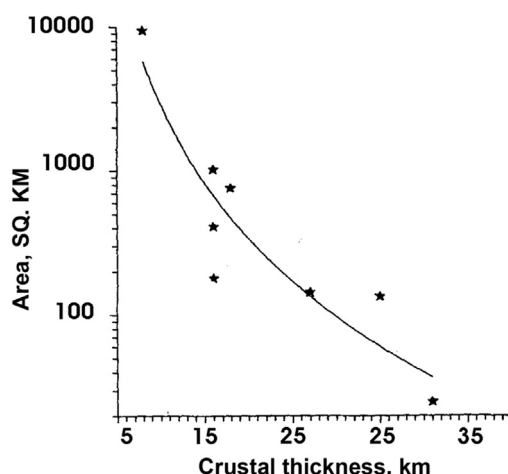


**FIGURE 8.4** Relationship between the epicentral area of volcano-tectonic earthquake swarms and the crustal thickness for eight selected eruptions. From Zobin, V.M., 1996. *The epicentral areas of volcanic earthquakes and their volcanotectonic significance. Geofísica Internacional* 35, 95–103.

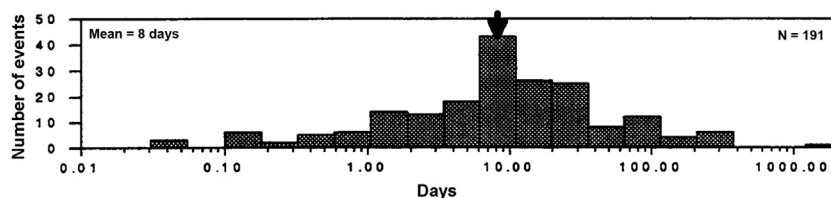


**FIGURE 8.5** Histogram of the duration of 191 volcanic earthquake swarms preceding eruptive activity all over the world. The data form a nearly log-normal distribution with a geometric mean of 8 days. From Benoit, J.P., McNutt, S.R., 1996. *Global volcanic earthquake swarm database and preliminary analysis of volcanic earthquake swarm duration. Annals of Geophysics* 39, 221–229.

## 8.2.2 Earthquake Swarm Duration

Benoit and McNutt (1996) have compiled a Global Volcanic Earthquake Swarm Database, 1979–89. The database gives the data about 191 swarms preceding eruption activity, as reported in the *Bulletin of Volcanic Eruptions* of the Volcanological Society of Japan (Fig. 8.5).

Fig. 8.5 shows that the main part of earthquake swarms preceding eruptive activity had durations from 1 to 100 days (or from 24 to 2400 hours). A geometric mean of preliminary swarm duration is 8 days (132 hours). This value is slightly longer than the value ( $117 \pm 7$  hours) obtained from the data in Table 8.1.

Benoit and McNutt (1996) made an interesting observation about the relationship between the swarm duration and eruption repose time. Fig. 8.6 shows that the long swarms (longer than 4 months) have a good correlation

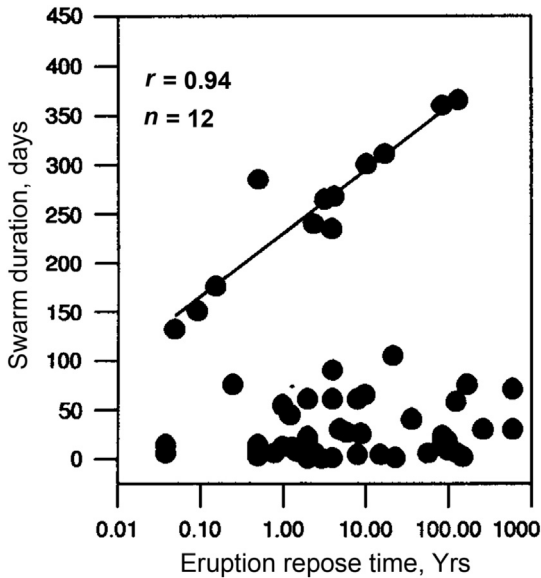


**FIGURE 8.6** Relationship between the duration of 65 precursory swarms of the volcanoes all over the world and corresponding eruption repose time. The regression line shown is for 12 swarms with a duration of 100 days or greater. The correlation coefficient is 0.94, which is significant at the 99% confidence level. *From Benoit, J.P., McNutt, S.R., 1996. Global volcanic earthquake swarm database and preliminary analysis of volcanic earthquake swarm duration. Annals of Geophysics 39, 221–229.*

of their duration with the time of repose; the longer the period between eruptions, the greater the duration of seismic activity observed before the eruption. The authors assumed that longer-lasting swarms may be the result of ascending magma encountering a cooler and stronger crustal environment.

### 8.2.3 Magnitude–Frequency Relations of Events in Volcano-Tectonic Earthquake Swarms

The frequency of the occurrence of earthquakes in a given region and for a given period of time can be represented by the equation  $\log N = a - bM$ , where  $N$  is a number of earthquakes with magnitudes of a fixed range around magnitude  $M$ , and  $a$  and  $b$  are constants. The constant  $b$  is called the  $b$ -value and depends on the relation between large and small events in the earthquake sequence. For tectonic earthquakes,  $b$ -values are between  $2/3$  and  $1$  and do not show much regional variability.

For volcanic earthquake swarms,  $b$ -values were estimated as close to  $1$  or greater, up to  $2.6$  (Fig. 8.7), indicating that volcano-tectonic earthquake swarms tend to have an anomalously large number of small events compared to the usual seismic sequence (Gresta and Patané, 1983; McNutt, 1986;

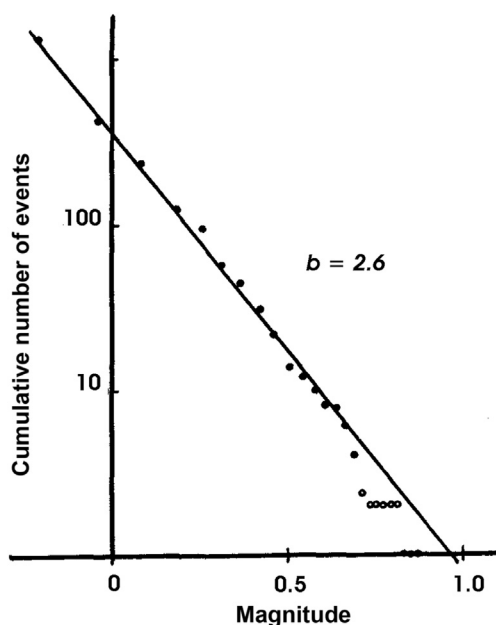


**FIGURE 8.7** A log-linear magnitude–frequency plot for 1316 B-type volcano-tectonic earthquakes recorded from December 1, 1974, to December 8, 1974, during the warning stages of the period of explosive eruptions at Pavlof volcano, Alaska. This plot was approximated with the relationship  $\log N = 2.53 - 2.6 M$ . From McNutt, S.R., 1986. *Observations and analysis of B-type earthquakes, explosions, and volcanic tremor at Pavlov volcano, Alaska. Bulletin of the Seismological Society of America* 76, 153–175.

Zobin et al., 2002). Results of McNutt (1986) for Pavlof volcano, Alaska show that  $b$ -values of volcano-tectonic microearthquake ( $M = 0-1$ ) swarms do not depend on the stages of volcanic activity.

At the same time, when larger earthquakes participate in the swarm, the magnitude–frequency distribution can vary during the volcanic activity or be non–log-linear (Gresta and Patané, 1983; Okada, 1982). This reflects the unstable relationship between small and large events during the volcano-tectonic earthquake swarm. Gorshkov (1959) has noted that during the course of an earthquake swarm related to the 1956 eruption of Bezymianny volcano, Kamchatka (see Section 6.1.1), there were observed two independent sequences of large and small events. The large events (magnitude about 4) occurred as a background without any relation to the course of the eruption, while the smaller events reflected the variations in volcanic activity. Okada (1982) demonstrated the same characteristics for the earthquake swarm that preceded the 1980 cataclysmic eruption of Mount St. Helens, Cascades volcano (see Section 7.1.1). Fig. 8.8 shows that, in the case of Mount St. Helens, the large and small earthquakes developed in time distinctively. There exists an intrinsic lack of earthquakes over a particular magnitude range.

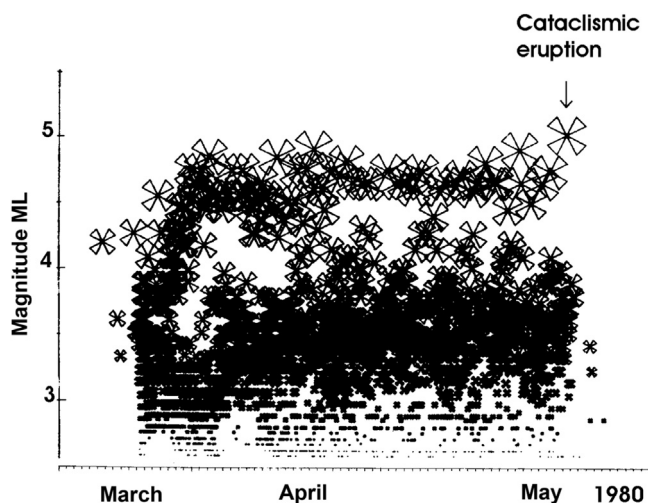


**FIGURE 8.8** Temporal variations in magnitudes of volcano-tectonic earthquakes preceding the May 18, 1980 directed blast at Mount St. Helens volcano, Cascades. From Okada, H., 1983. *Comparative study of earthquake swarms associated with major volcanic activities*. In: Shimozuru, D., Yokoyama, I., (Eds.), *Arc Volcanism: Physics and Tectonics*, Terra Publ. Co., Tokyo, pp. 43–61.

The effect of the intrinsic lack of earthquakes within the volcano-tectonic earthquake swarms may be a reason for the non-log-linear magnitude–frequency relations. Okada (1982) pointed out this effect for three volcano-tectonic earthquake swarms associated with the great eruptions of Fernandina, Galapagos (1968), Usu, Hokkaido (1977–78), and Mount St. Helens, Cascades (1980; Fig. 8.9). It is seen that for all three cases, there is a minimum in magnitude–frequency relations. Okada (1982) considers that larger earthquakes are related to a large amount of surface deformation in the form of doming, caldera collapse, or bulging. Therefore the volcano-tectonic earthquake swarm may consist of two or more separate swarms reflecting the different scale processes during great manifestations of volcanic activity.

### 8.3 SOME REGULARITIES IN THE VOLCANO-TECTONIC EARTHQUAKE SWARMS PROCLAIMING REAWAKENING OF ANDESITIC AND DACITIC VOLCANOES

The general cycle of activity of andesitic and dacitic volcanoes (Chapters 6 and 7) may be described as follows:

1. Quiescence.
2. Phreatic explosive activity with the formation of a new crater.
3. Repeated dome building and destruction accompanied by explosions.
4. Quiescence.

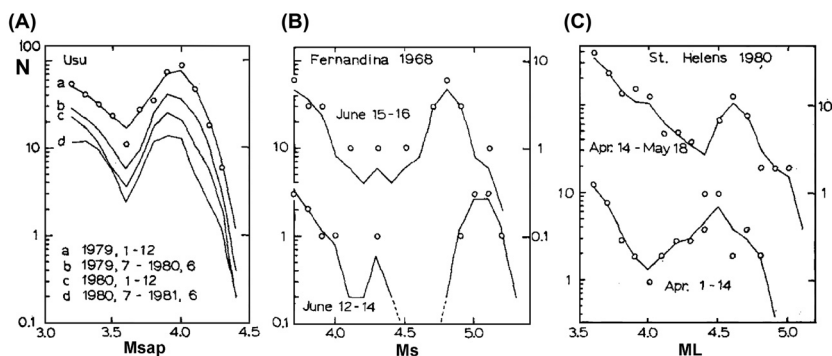


**FIGURE 8.9** Non-log-linear magnitude-frequency plots for volcano-tectonic earthquakes associated with the 1979–80 lava dome activity at Usu, Hokkaido (A), the 1968 caldera collapse at Fernandina, Galapagos (B), and the 1980 directed blast at Mount St. Helens, Cascades (C).  $N$  is the number of earthquakes;  $Msap$  is the magnitude determined from Sapporo station records. The plots are shown for different stages of volcanic activity indicated by dates. From Okada, H., 1983. *Comparative study of earthquake swarms associated with major volcanic activities*. In: Shimozuru, D., Yokoyama, I., (Eds.), *Arc Volcanism: Physics and Tectonics*, Terra Publ. Co., Tokyo, pp. 43–61.

The period of quiescence may continue for a few years or a few centuries. Usually, the largest explosive eruptions occur when the period of quiescence is at least 100 years. Among them are the explosions at volcanoes Bezymianny (1956, Kamchatka), Mount St. Helens (1980, Cascades), El Chichón (1982, Mexico), and Pinatubo (1991, Philippines). All these large eruptions were preceded by swarms of volcano-tectonic earthquakes described in Chapters 6 and 7. The volcano-tectonic earthquakes followed the main stages of the volcanic eruptions and may serve as a good tool for reconstruction of the stress release during eruptive events.

Six significant eruptions at volcanoes with a long quiescence that occurred during last 50 years were selected (Table 8.2): Bezymianny (Kamchatka, 1956), Mount St. Helens (Cascades, 1980), El Chichón (Mexico, 1982), Unzen (Japan, 1991), Pinatubo (Philippines, 1991), and Soufrière Hills (Montserrat, West Indies, 1995). The main criteria for the selection of this group of volcanoes were as follows: a long period of quiescence (more than 100 years) and the presence of seismic data associated with the different stages of their activity. The selected eruptions were characterized by different values of volcanic explosivity index (VEI, see Chapter 1: for VEI definition), ranging from VEI 1 for the eruptions of Unzen in 1991 to VEI 6 for Pinatubo in 1992.

Fig. 8.10 illustrates the development of seismic activity during the reawakening of the Bezymianny volcano during 1955–56 (see Section 6.1.1). It is possible to discriminate three stages in the reawakening process: (1) the

**TABLE 8.2** Characteristics of Volcano-Awakening Process at Andesitic and Dacitic Volcanoes.

| Volcano                                   | VEI | Time of Quiescence (years) | Duration of Stage 1 (Seismic Swarm Before the First Explosion, days) | Duration of Stage 2 (Before the Great or Large Explosion, days) |
|-------------------------------------------|-----|----------------------------|----------------------------------------------------------------------|-----------------------------------------------------------------|
| Bezymianny, Kamchatka, 1955–56            | 5   | 900–1000                   | 24                                                                   | 160                                                             |
| Mount St. Helens, Cascades, 1980          | 5   | 120                        | 11                                                                   | 52                                                              |
| El Chichón, Mexico, 1982                  | 5   | 600                        | 28                                                                   | 7                                                               |
| Unzen, Kyushu, 1991                       | 1   | 200                        | 360                                                                  | 87                                                              |
| Pinatubo, Philippines, 1991               | 6   | 500                        | 20                                                                   | 70                                                              |
| Soufrière Hills volcano, Montserrat, 1995 | 3   | 370                        | 180–250                                                              | 33                                                              |

*Notes:* The values of VEI and the time of quiescence are given according to the Global Volcanism Program catalog (<http://www.volcano.si.edu/>, 31 May 2006).

*Source:* The durations of stages are taken from Zobin, V.M., Jiménez, Z., 2008. Some regularity in the process of re-awakening of andesite and dacite volcanoes: specific features of the 1982 El Chichón volcano, México reactivation. *Journal of Volcanology and Geothermal Research* 175, 482–487.

preliminary stage that was characterized by a sharp increase in seismic activity near the volcano up to the first phreatic explosion of October 22, 1955; (2) the stage between the first phreatic explosion and the large explosion of March 31, 1956, characterized by a low in number but high in magnitude seismic events; and (3) the postexplosion stage of dome building with a gradual decrease in the number and magnitude of seismic events.

The three stages are common for reawakening volcanoes (see Chapters 6 and 7) and maybe a basis for the analysis of any peculiarities in the awakening process. Table 8.2 summarizes the characteristics of the three stages of seismovolcanic activity for the six volcanoes studied. Relationships between the durations of Stages 1 and 2 and the VEI of the forthcoming large explosion are illustrated in Fig. 8.11.

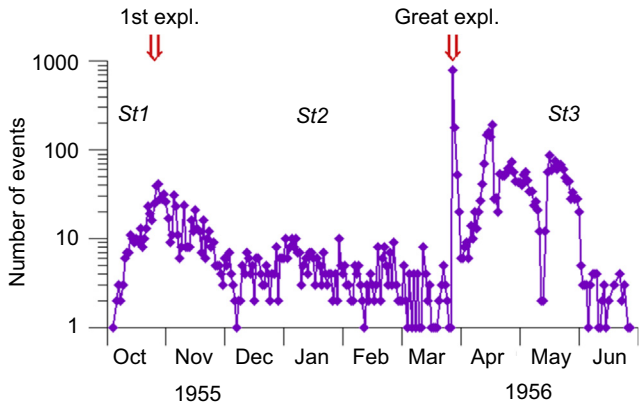


**FIGURE 8.10** Daily distribution of the number of earthquakes associated with the 1955–56 eruption of Bezymianny volcano. St 1, St 2, and St 3 indicate the three stages in the development of the activity. Arrows show the beginning of the eruption and the appearance of the great directed explosion of March 30, 1956. Data from Tokarev, P.I., 1981. *Volcanic Earthquakes of Kamchatka*. Nauka, Moscow (in Russian). 164 p.

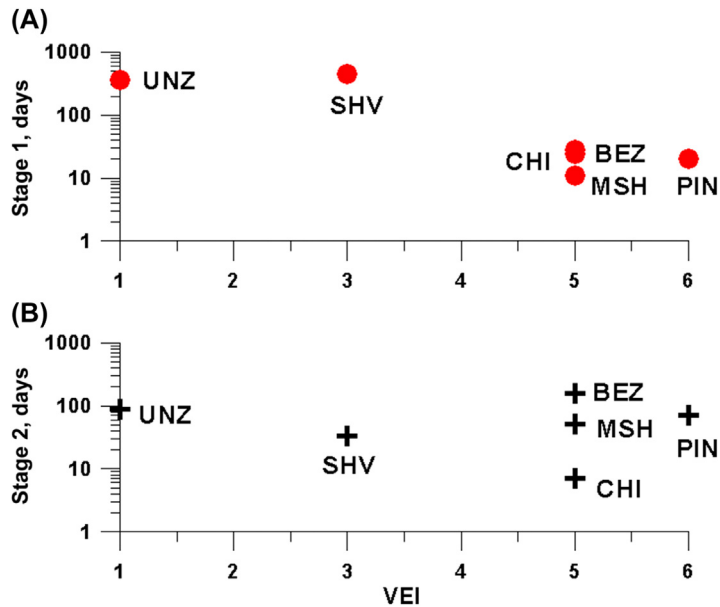


**FIGURE 8.11** Relationships between the duration of Stage 1 (A), and Stage 2 (B), and the size of VEI. The volcano indices are: UNZ, Unzen; SHV, Soufrière Hills volcano; CHI, El Chichon; BEZ, Bezymianny; MSH, Mount St. Helens; PIN, Pinatubo. From Zobin, V.M., and Jiménez, Z., 2008. Some regularity in the process of re-awakening of andesite and dacite volcanoes: specific features of the 1982 El Chichón volcano, México reactivation. *J. Volcanol. Geotherm. Res.* 175, 482–487.



### 8.3.1 Relationship Between the Duration of Stage 1 and the VEI of Forthcoming Explosion

Two groups of events are discriminated clearly in Fig. 8.11A: the group of low VEI and long duration Stage 1 events (Unzen, 1991 and Soufrière Hills volcano, 1995) and the group of high VEI and short duration Stage 1 (Bezymianny, 1956; Mount St. Helens, 1980; El Chichón, 1982; and Pinatubo, 1991). This discrimination shows that long-lasting seismic activity (about or more than 1 year) before the reawakening of a volcano signifies a low probability that the volcano will produce a large explosion. On the contrary, rapid development (less than a month) of seismic activity may proclaim a forthcoming large-magnitude explosion.

### 8.3.2 Relationship Between the Duration of Stage 2 and Postexplosion Dome Building

Generally, it is impossible to show any relationship between the duration of Stage 2 from the first phreatic explosion at a reawakening volcano to its largest explosion and VEI value (Fig. 8.11B). One significant exception that should be noted is that the only volcano whose sequence of explosions did not culminate in a dome-building period was El Chichón. Its 1982 eruption was very short, and its Stage 2 was for 7 days only. It is the lowest value for the duration of Stage 2 in Fig. 8.11B. All other eruptions, whose Stage 2 continued for more than 20 days, produced significant lava domes.

### 8.3.3 Conceptual Model of Reawakening Process

Zobin and Jiménez (2008) proposed the conceptual model for the three stages in the reawakening of dormant volcanoes based on models of lava dome formation (McLeod and Tait, 1999; Melnik and Sparks, 2002).

Stage 1 is the first announcement of the forthcoming eruption of a dormant volcano. Therefore it may be considered as a reflection of the process of deep dyke nucleation. Considering dyke growth following chamber pressurization, two alternatives may be proposed (McLeod and Tait, 1999): sudden chamber pressurization or gradual chamber pressurization. Dyke growth that is controlled by sudden chamber pressurization may be more rapid than that controlled by gradual chamber pressurization.

Seismic activity is indicative of a passageway for magma to reach the surface and would be a function of the type of dyke growth. In this case, we may consider the slow development of seismic activity as a result of gradual chamber pressurization. On the contrary, the rapid development of seismic activity may be considered as a result of sudden chamber pressurization. In both cases, decompressive degassing of the intruding magma during final seismic activity leads to shallow pressurization that produces the first explosion (Gardner and White, 2002).

Stage 2 may be attributed, in the frame of the model by Melnik and Sparks (2002), to the nonlinear effects of crystallization and gas loss by permeable flow within magma. They note that degassing results in rheological stiffening due to gas exsolution and microlite crystallization from an under-cooled melt. Viscosity can increase by several orders of magnitude when silicic melts, containing a few percent of dissolved water, decompress, and degas. This stage is characterized by a process of gas exsolution from the melt that can produce a sequence of ash explosions forming a new crater. Only a small portion of the fresh magma material is expelled. A wide range of Stage 2 durations was observed for different eruptions, from 7 to 160 days (Table 8.2). The duration of the eruption development from the first small phreatic explosion to a large phreatic or phreatomagmatic explosion may be proportional to the rate of gas exsolution and dissolved water liberation from magma melt. The intensity of a large explosion is supposed to depend on the type of magma chamber pressurization (or on the duration of Stage 1): Sudden magma chamber pressurization leads to a very large explosion whereas gradual magma chamber pressurization leads to only a large explosion.

Stage 3 is characterized by an important volcanological aspect, whether it follows the construction of a new lava dome or not. This aspect is supposed to be constrained by the duration of Stage 2. The only case where no dome was built was the eruption at El Chichón volcano, and this eruption was characterized by the shortest duration of Stage 2. Another feature of this eruption was the high intensity of all the explosions. Considering that the viscosity of silicic magma containing only a few percent of dissolved water can increase by several orders of magnitude during decompression and degassing (Melnik and Sparks, 2002), it may be supposed that a high rate of decompressing and degassing during a short time had increased the viscosity within the dyke up to a level when it was impossible to push it to the surface. As a result, a dyke was formed as a degassed intrusive structure beneath the volcano without building a superficial lava dome.

It is important to consider the three separate stages during the development of a strategy of alarms when a volcano begins to awake. The first stage is the monitoring the awakening process. The second stage is important for understanding that the first explosion of the awakening volcano is not the strongest. The strongest one will occur after some time.

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- Zobin, V.M., Jiménez, Z., 2008. Some regularity in the process of re-awakening of andesite and dacite volcanoes: specific features of the 1982 El Chichón volcano, México reactivation. *Journal of Volcanology and Geothermal Research* 175, 482–487.

## RELEVANT WEBSITES

- BGVN. Bull. Glob. Volc. Network. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.
- GVP. Global Volcanic Program. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.
- Volcanological Society of Japan <http://www.kazan-g.sakura.ne.jp/E/index.html>.

## Chapter 9

# Source Properties of Volcano-Tectonic Earthquakes

In this chapter, the author will discuss the focal mechanisms and source properties of volcano-tectonic earthquakes. In Chapters 5–8, it was shown that volcano-tectonic earthquakes occur within and around the volcanic edifice and reflect the interaction of two general geological processes: the magma migration to the Earth's surface and the crustal tectonic activity. Therefore it is important to study the focal mechanisms of volcano-tectonic earthquakes and stress release during volcano-tectonic earthquake swarms, and the relationship between the stress situation within the volcanic edifice and the regional stress field.

### 9.1 FOCAL MECHANISMS OF VOLCANO-TECTONIC EARTHQUAKES: DOUBLE-COUPLE AND NON-DOUBLE-COUPLE MODELS

The complete tensor of a volcano-tectonic earthquake source may be described as a sum of its isotropic and deviatoric parts (see Section 4.3). The deviatoric part may be decomposed into two source types: a double couple (DC) and a compensated linear vector dipole (CLVD). The application of these two source types to describe the fault nature of volcano-tectonic earthquakes is discussed.

#### 9.1.1 Double-Couple Model

During the period 1962–64, the global seismic networks were installed. It allowed collecting enough *P*-wave first motions to construct the focal mechanism solution on the stereographic projection for volcanic earthquakes of magnitude about 5 or greater. The first *P*-wave motion focal mechanism solutions for volcano-tectonic earthquakes (Fig. 9.1) were constructed for two magnitude 5 earthquakes preceding the November 11, 1964 directed-blast eruption at Sheveluch volcano, Kamchatka (Zobin, 1970). They indicated a strike-slip (SS) faulting for these earthquakes. Zobin (1972, 1979a,b) constructed the focal mechanism solutions for the magnitude 5–6 volcano-tectonic earthquakes associated with the eruptions of Miyakejima, Tori-shima, Beerenberg,

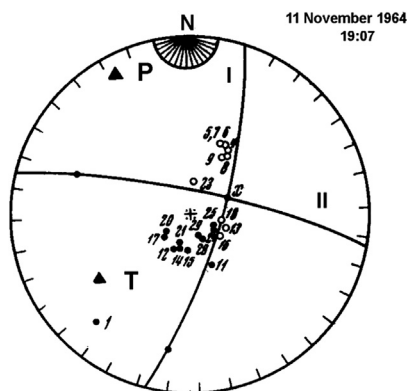


**FIGURE 9.1** Focal mechanism solution (upper hemisphere projection) for the volcano-tectonic earthquake that coincided with the 11 November 1964 directed blast at Sheveluch volcano, Kamchatka. I and II are possible fault planes; *P* is compression; *T* is tension; solid circles, compressions in P-wave arrivals; open circles, dilatations in P-wave arrivals. Numbers indicate the seismic stations. From Zobin, V.M., 1970. *Focal mechanism of volcanic earthquakes related to the Sheveluch volcano eruption in November 1964*. In: *Izvestia, Solid Earth Physics, USSR Acad. Sci*, vol. 3, pp. 31–36 (in Russian). A.G.U. Transl. pp. 161–164.

and New Tolbachik volcanoes. All of them were of SS faults. It was shown that one of two nodal planes of the focal mechanism solution of large volcano-tectonic earthquake coincides with the strike of the fissure, where new volcanic centers form (New Tolbachik volcanoes; Fig. 9.2). Zobin (1972) showed the similarity between focal faulting of these volcano-tectonic earthquakes and that of tectonic earthquakes surrounding the volcanic regions.

These first results were supported by studies of focal mechanisms that were developed for volcano-tectonic earthquakes associated with different eruptions in the world. Fig. 9.3 shows the focal mechanism solutions that were constructed for some volcano-tectonic earthquakes originating beneath the dacitic Mount St. Helens volcano, Cascades (Weaver et al., 1981) after the directed-blast eruption of May 18, 1980, during the stage of magmatic eruptions (see Section 7.1.1). These earthquake focal mechanisms indicated the thrusts (A–C and E) and SS faulting at the volcano-tectonic earthquake foci. All focal mechanisms have steeply dipping fault planes that strike northeast.

The study of the focal mechanisms of earthquakes, that occurred during the same year (1980) within the tectonic structures to the north and south of the Mount St. Helens volcano (Fig. 9.4), shows the similarity in focal mechanisms of the earthquakes, originated beneath the volcano at the active eruptive stage, and the earthquakes, originating in tectonic ranges far from the volcano. They were the same thrusts (indexes F and I in) and SS (G, J, K, and L), as well as normal faults (H).

Focal mechanism solutions for volcano-tectonic earthquakes preceding the 1989 Teishi Knoll submarine basaltic eruption (see Section 5.5.1) were obtained

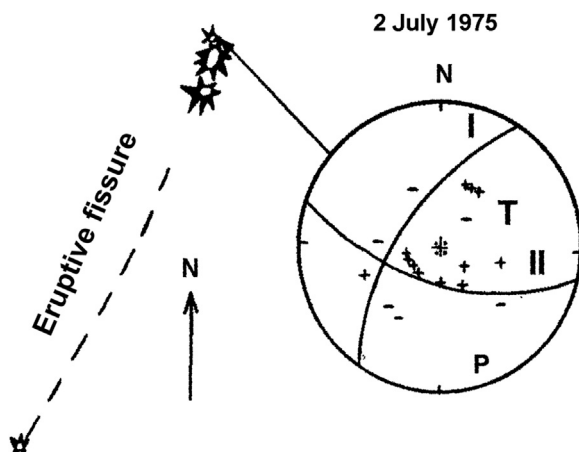


**FIGURE 9.2** Focal mechanism solution (lower hemisphere projection) for the 2 July (04:07) volcano-tectonic earthquake with magnitude,  $M_w 4.9$  preceding the 6 July 1975 New Tolbachik fissure eruption, Kamchatka. It is seen that the fault plane I of the solution coincides with the direction of the 12-km length fissure where the chain of new cones formed. I, II are the possible fault planes; P is compression; T is tension; symbol (+) shows compressions in P-wave arrivals; symbol (-) shows dilatations in P-wave arrivals. From Gorelchik, V.I., Zobin, V.M., Tokarev, P. I., 1990. Volcanic earthquakes of Kamchatka: classification, nature of source and spatio-temporal distribution. *Tectonophysics* 180, 255–271.

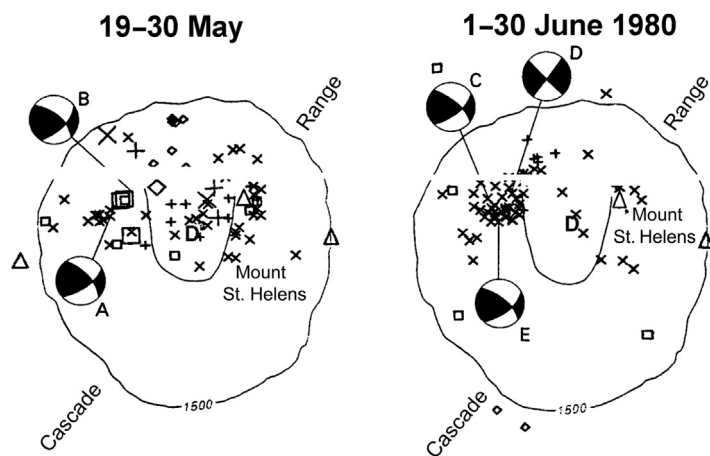


**FIGURE 9.3** Focal mechanism solutions of volcano-tectonic earthquakes associated with the 1980 magmatic activity of Mount St. Helens volcano, Cascades and the epicenters of these earthquakes. Epicenters of earthquakes are shown as small symbols (magnitudes less than 2.8) and large symbols (magnitudes greater than 2.8). Depths are indicated with symbols as + (0–5 km), × (5–10 km), □ (10–15 km) and ◇ (greater than 15 km). Post-eruption 1500-m contour is shown. The location of the lava dome is shown as D. The focal mechanism diagrams are upper hemisphere plots, with darkened quadrants indicating compression and white quadrants indicating dilatations; solid circles, compressions; open circles, dilatations. Mechanism B is a composite solution; all other solutions are single-event solutions. From Weaver, C.S., Grant, W.C., Malone, S.D., Endo, E.T., 1981. Post-May 18 seismicity: volcanic and tectonic implications. In: Lipman, P.W., Mullineaux, D.R. (Eds.), *The 1980 Eruptions of Mount St. Helens, Washington. Geological Survey Professional Paper 1250*, pp. 109–121.

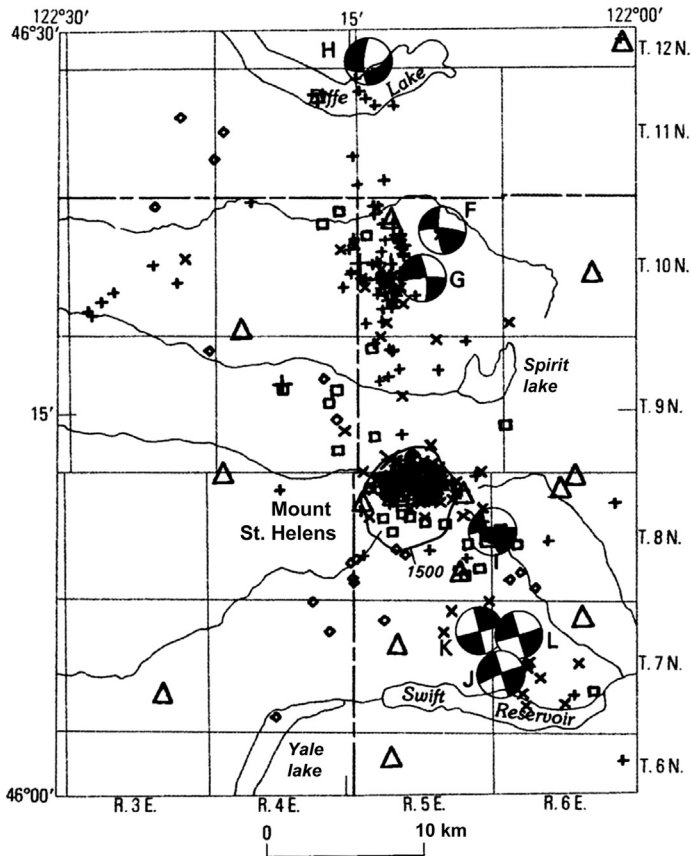


**FIGURE 9.4** Seismicity and focal mechanism diagrams for some events that occurred within the tectonic environment of Mount St. Helens between 18 May and 31 August 1980. Legend is the same as for Fig. 9.3. From Weaver, C.S., Grant, W.C., Malone, S.D., Endo, E.T., 1981. Post-May 18 seismicity: volcanic and tectonic implications. In: Lipman, P.W., Mullineaux, D.R. (Eds.), *The 1980 Eruptions of Mount St. Helens*, Washington. Geological Survey Professional Paper 1250, pp. 109–121.

by two methods: from *P*-wave polarity distribution (Matsumura et al., 1991) and from the inversion method (Takeo, 1992). Fig. 9.5 shows the *P*-wave first motion focal mechanism solutions for three earthquakes (Matsumura et al., 1991). The largest earthquake (magnitude JMA 5.5) revealed an SS nature with a normal fault component; two magnitude 3.1 earthquakes indicated thrust faulting.

The solution obtained by the inversion method (Takeo, 1992) for the same magnitude 5.5 earthquake is consistent with the *P*-wave first motion solution. Fig. 9.6 shows the source time functions of the SS and dip-slip components on each subfault of the earthquake fault model marked by arrows. It is seen that a right-lateral SS motion dominated during the rupturing.

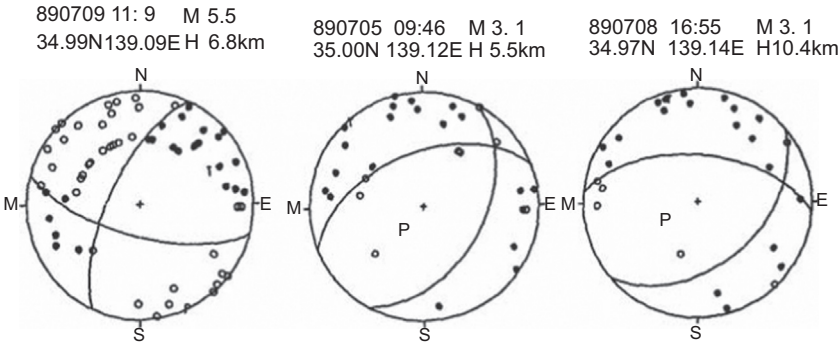


**FIGURE 9.5** Focal mechanism solutions (lower hemisphere projection) for three volcano-tectonic earthquakes preceding the 13 July 1989 submarine Teishi Knoll eruption, Izu Islands. *P* is compression; *T* is tension; *solid circles*, compressions in *P*-wave arrivals; *open circles*, dilations in *P*-wave arrivals. From Matsumura, S., Ohkubo, T., Imoto, M., 1991. Seismic swarm activity in and around the Izu Peninsula preceding the volcanic eruption of July 13, 1989. *Journal of Physics of the Earth* 39, 93–106.

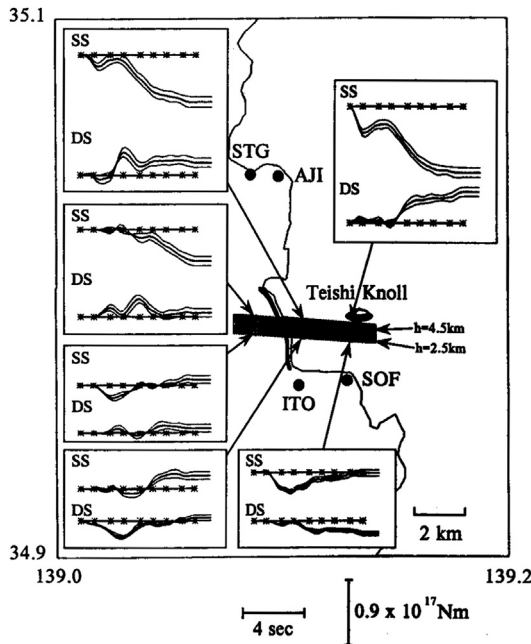


**FIGURE 9.6** Spatial variation of source time functions with strike-slip (SS) and dip-slip (DS) components for the magnitude 5.5 (July 9, 1989) earthquake preceding the July 13, 1989 submarine Teishi Knoll eruption, Izu Islands. The seismic stations whose records were used in the inversion analyses are shown by *filled circles*. The *arrows* show the location of six point sources, which represents a 2 km  $\times$  2 km subfault. From Takeo, M., 1992. The rupture process of the 1989 offshore Ito earthquakes preceding a submarine volcanic eruption. *Journal of Geophysical Research* 97, 6613–6627.



The focal mechanism solutions that were determined by the *US National Earthquake Information Center* (NEIC) for seven magnitude  $m_b$  4.9–5.7 earthquakes, which occurred during the paroxysmal stage of the eruption during June 12, 1991–June 15, 1991 of the dacitic volcano Pinatubo, Philippines (see Section 7.1.4), revealed the predominantly SS movement with a minor vertical (normal or reverse) component; only one earthquake was from normal faulting with a large SS component. These solutions suggest that the largest volcano-tectonic earthquakes at Pinatubo were consistent with the east–west maximum regional stress (Bautista et al., 1996).

Okada et al. (1981) determined the composed focal mechanism solutions for small volcano-tectonic earthquakes, which occurred at Usu volcano, Hokkaido during the different stages of dacitic dome activity from October 1977 to June 1979. Dip-slip type of faulting with slight variations within the summit zone was dominant. The strike of the steeply dipping fault planes was roughly parallel to the U-shaped major fault zone within the crater.

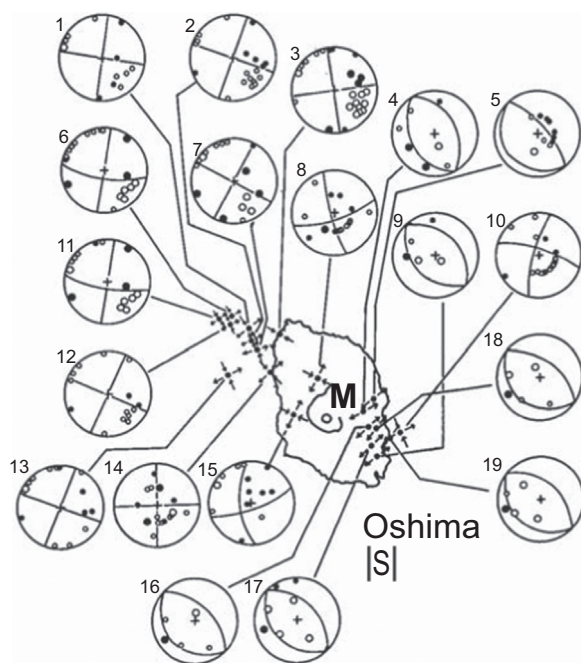


**FIGURE 9.7** Focal mechanism solutions of volcano-tectonic earthquakes that occurred before and after the 1986 fissure eruption of Oshima, Izu volcano. *M* marks the summit crater Miharayama; solid circles, compressions; open circles, dilatations. Arrows show the *P* and *T* axes. From Yamaoka, K., Watanabe, H., Sakashita, S., 1988. Seismicity during the 1986 eruption of Izu-Oshima volcano. *Bulletin of the Volcanological Society of Japan* 33, S91–S101.

During the 1986–87 activity of basaltic volcano Oshima, Izu islands, focal mechanism solutions were determined for earthquakes that occurred before and after the fissure eruption. A zone to the north of the volcano was characterized by right-lateral SS faults, whereas a zone to the south of the volcano was characterized by normal faults (Fig. 9.7). The SS fault motions within the first zone are consistent with the common tectonic stress field around Oshima Island; the peculiar normal fault motions at the second zone may reflect triggering by the fissure eruptions in the tensional stress field (Yamaoka et al., 1988).

Therefore the fault nature of volcano-tectonic earthquakes may be thrust, normal fault, or SS. It may be constrained by the regional or local tectonic stress systems. The focal mechanisms of relatively large earthquakes (magnitude about 5–6), related to fissure eruptions, revealed the coincidence of its one fault plane with the strike of the eruptive fissure.

### 9.1.2 Non-Double-Couple Model

The non-DC sources, or the sources with values  $0.2 \leq \varepsilon \leq 0.5$  ( $\varepsilon$  is the relationship between the smallest and largest eigenvalues of the seismic moment

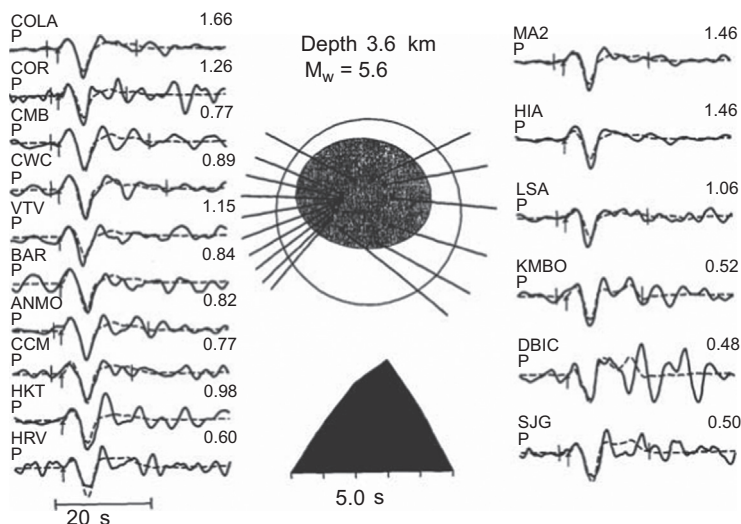


**FIGURE 9.8** Comparison of the calculated synthetic seismograms (dashed) with the broadband teleseismic P-waveforms (solid) for the September 29, 1996 volcano-tectonic earthquake preceding the September 30 flank eruption of Grimsvötn volcano, Iceland. The focal mechanism corresponds to the full moment tensor solution, and the source time function is determined by the inversion. Maximum amplitudes (in microns) are shown to the right of the seismograms, *arrows* show picked first arrivals, and the *thin vertical bars* show the time window included in the inversion. From Nettles, M., Ekström, G., 1998. Faulting mechanism of anomalous earthquakes near Bárðabunga volcano, Iceland. *Journal of Geophysical Research* 103, 17973–17983.

tensor), were determined for a few volcano-tectonic earthquakes located at ring fault structures in Iceland (Ekström, 1994; Nettles and Ekström, 1998) and at Etna and Vesuvius, Italy (Panza and Saraó, 2000; Saraó et al., 2001). Fig. 9.8 shows an example of the focal mechanism determined for the September 29  $M_w$  5.6 main shock of earthquake sequence preceding the September 30, 1996 eruption of the subglacial volcano Grimsvötn, Iceland, using broadband teleseismic  $P$ -waveforms inversion. The  $\varepsilon$  value for this moment tensor solution was equal to 0.30. Similar solutions were obtained for nine earthquakes that occurred from 1976 to 1994 around the Bárðarbunga caldera (Fig. 9.9). The  $\varepsilon$  values for these earthquakes vary from 0.24 to 0.41, and the magnitudes  $M_w$  range from 5.1 to 5.4. They occurred within the range of depths from 3.3 to 6.7 km. Nettles and Ekström (1998) consider that these non-DC solutions may reflect the faulting on an outward dipping cone-shaped ring fault, beneath the Bárðarbunga caldera. The model of rupture propagation on a cone-shaped fault is shown in Fig. 9.10 and explains the rupturing as a series of subevents propagating around the surface of the cone.

Saraó et al. (2001) studied the variations in the percentage of DC and CLVD components in the sources of volcano-tectonic earthquakes preceding the 1991 eruption at Etna, Sicily (see Section 5.1.2). Fig. 9.11 shows that between August 1990 and December 9, 1991, the percentage of non-DC components of earthquakes, statistically significant at the 95% confidence level, increased

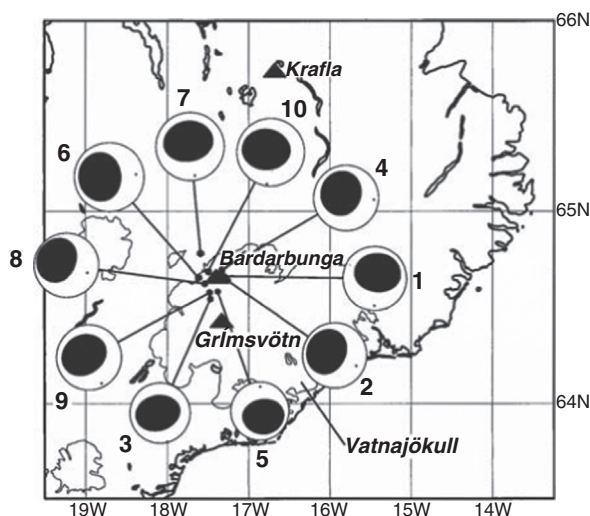


**FIGURE 9.9** Map showing Harvard CMT solutions for 10 earthquakes that occurred near Bárðarbunga volcano, Iceland. The earthquakes are highly non-double couple, and have vertical  $T$ -axes, indicative of horizontal compression. From Nettles, M., Ekström, G., 1998. Faulting mechanism of anomalous earthquakes near Bárðarbunga volcano, Iceland. *Journal of Geophysical Research* 103, 17973–17983.

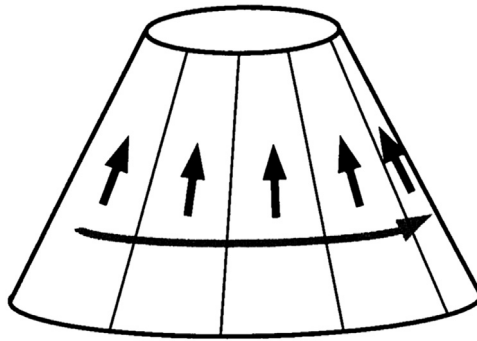


**FIGURE 9.10** Model of rupture propagation on a cone-shaped fault : a subsurface with reverse-faulting geometry (*short arrows*) occurs on each fault segment in succession (*long arrow*), around a specified portion of the cone. From Nettles, M., Ekström, G., 1998. Faulting mechanism of anomalous earthquakes near Bárðabunga volcano, Iceland. *Journal of Geophysical Research* 103, 17973–17983.

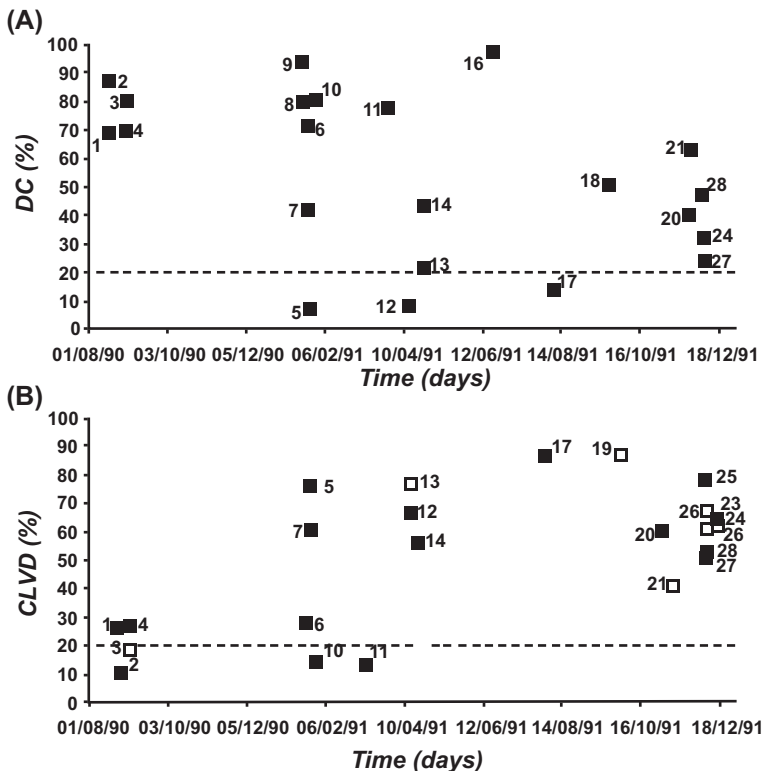


**FIGURE 9.11** Temporal variation of the percentage of DC (A) and CLVD (B) components of the earthquakes preceding the December 15, 1991 flank eruption at Etna, Sicily. The *dashed lines* represent the limits below which the solutions may be due to numerical noise. Numbering corresponds to temporal appearance of earthquakes. DC, double couple; CLVD, compensated linear vector dipole. From Saraó, A., Panza, G.F., Privitera E., Cocina, O., 2001. Non-double-couple mechanisms in the seismicity preceding the 1991–1992 Etna volcano eruption. *Geophysical Journal International* 145, 319–335.

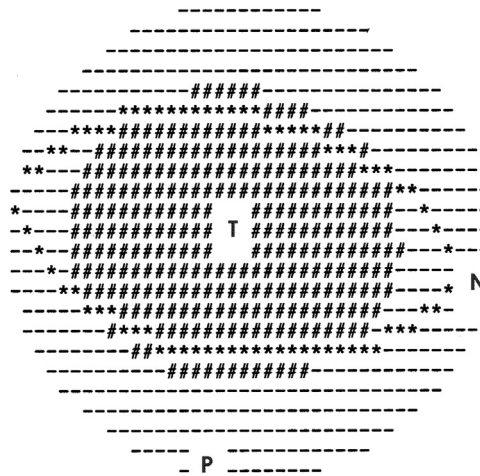


**FIGURE 9.12** Standard Harvard CMT solution showing two possible interpretations of the solution for the September 29, 1996 Grimsvötn earthquake : a double-couple source (marked by stars) or a non–double-couple source (marked by filled ellipse area) T, tension; P, pressure; N, neutral axes for the best double couple. From Harvard CMT, <http://www.globalcmt.org/CMTsearch.html>.

before the beginning of the December 15, 1991 flank eruption. Panza and Saraó (2000) consider that this fact can be related to fluid movements.

Therefore the non-DC nature of the volcano-tectonic earthquake source may reflect the ring (or other complex) structure of the fault generating the earthquake (Nettles and Ekström, 1998). At the same time, every complex fault may be approximated by a linear structure, and for the majority of the events with the non-DC solution is obtained an adequate DC solution (Fig. 9.12).

The presence of non-DC components can also depict a process in which a depressurizing magma body and crack formation leads to magma injection toward the shallower portions of the volcano (Panza and Saraó, 2000).

## 9.2 SOURCE SPECTRAL CHARACTERISTICS OF VOLCANO-TECTONIC EARTHQUAKES

The spectra of volcanic earthquakes are used generally in two types of applications: the total spectra of seismic records may be used for the classification and monitoring of seismic activity, and the body-wave (or surface-wave) spectra may be used for the study of source parameters of earthquakes.

### 9.2.1 Spectra of Total Records of Volcano-Tectonic Earthquakes

Fig. 9.13 shows the seismograms and spectrograms of records for the two main types of volcano-tectonic earthquakes recorded during the active period at

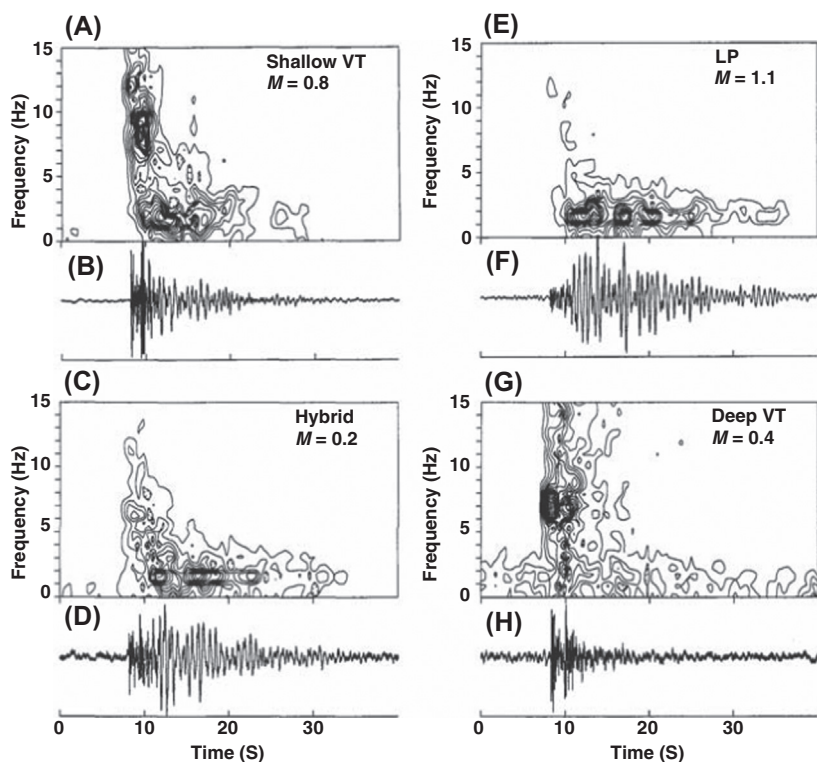


**FIGURE 9.13** Spectrograms (A, C, E, and G) and seismograms (B, D, F, and H) for earthquakes recorded at Redoubt volcano, Alaska during the 1989–90 activity. LP are the long-period earthquakes; deep VT are the short-period volcano-tectonic events that occurred at depths greater than 5 km below the crater floor; shallow VT are the short-period volcano-tectonic events that occurred at depths less than 5 km below the crater floor; and Hybrid are the hybrid earthquakes. From Lahr, J.C., Chouet, B.A., Stephens, C.D., Power J.A., Page, R.A., 1994. Earthquake classification, location, and error analysis in a volcanic environment: implications for the magmatic system of the 1989–1990 eruptions at Redoubt volcano, Alaska. *Journal of Volcanology and Geothermal Research* 62, 137–152.

Redoubt, Alaska during 1989–90 (Lahr et al., 1994). Each seismogram was of 40.6-s duration; the spectra were normalized for better comparison. The short-period events that occurred at depths greater than 5 km below the crater floor (deep volcano-tectonic in Fig. 9.13) are characterized by a peak between 6 and 8 Hz corresponding to the body-wave record and a low-frequency maximum between 1 and 3 Hz generated by short seismic coda. Shallower short-period events (shallow volcano-tectonic in Fig. 9.13) have two peaks: (1) a more intensive peak between 1 and 3 Hz and (2) a lower peak between 7 and 9 Hz. The first peak is produced by surface waves, while the second is produced by body waves.

The spectral contents of both types of volcano-tectonic earthquakes differ from the spectral contents of long-period and hybrid earthquakes also

recorded at Redoubt during the 1989–90 activity and shown in Fig. 9.13. The long-period events are characterized by a peak frequency near 1.5 Hz corresponding to S- and surface waves. A spectrum of a hybrid earthquake is characterized by a peak between 1 and 2 Hz. These differences in the spectral content of different types of volcano-tectonic earthquakes in comparison with other seismic events allow their visual discriminating for monitoring of volcano activity.

### 9.2.2 Spectral Source Characteristics of Volcano-Tectonic Earthquakes

The spectra of body and surface waves of volcano-tectonic earthquakes, as well as synthetic seismogram modeling, may be used for the estimation of earthquake source parameters, such as the seismic moment  $M_0$ , stress drop  $\Delta\sigma$ , and spectral corner frequency  $f_c$ . These parameters describe the variations of physical properties within the seismic volume below the volcano and can serve to discriminate between the seismic events generated during the different types of volcanic activity. In this chapter, the author used the estimations of source parameters from the Harvard centroid-moment tensor (Harvard CMT) and from body-wave spectra (circular dislocation Brune's model; Brune, 1970).

Brune's model (1970) allows the calculation of the radius of circular dislocation  $r$ , the seismic moment  $M_0$ , and stress drop  $\Delta\sigma$  from the S-wave spectrum using the following formulas:

$$M_0 = \frac{(4\pi\rho\beta^3 S_0)}{(G(d)C(\omega)R(\theta, \varphi))} \quad (9.1)$$

$$r = \frac{2.34\beta}{2\pi f_c} \quad (9.2)$$

$$\Delta\sigma = \frac{7M_0}{(16r^3)} \quad (9.3)$$

Here  $\pi = 3.14$ ;  $\rho$  is the assumed average density of the medium;  $\beta$  is the S-wave velocity;  $S_0$  is the spectral density at zero frequency;  $f_c$  is the corner frequency of the intersection of the low- and high-frequency branches of the spectrum;  $G(d)$  is the function of geometrical spreading and taken as the inverse hypocentral distance  $d$ ;  $C(\omega) = 2$  is the local transfer function; the S-wave radiation pattern  $R(\theta, \varphi)$  may be taken as 0.632 (the root mean square average of the shear displacement radiation pattern over entire focal sphere). The values of  $S_0$  and  $f_c$  may be taken from the spectra of the S-wave as shown in Fig. 9.14. This methodology may be also applied to the spectra of  $P$ -waves (Hanks and Wyss, 1972).



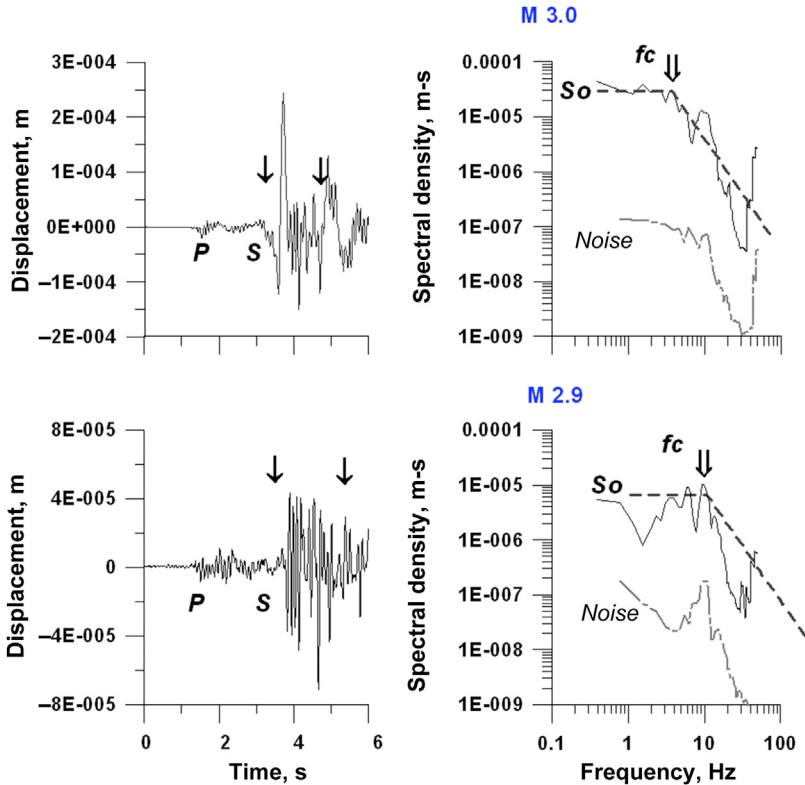


**FIGURE 9.14** Examples of S-waveforms and corresponding Fourier spectra of the 2000 Usu volcano-tectonic earthquakes recorded at station. NBBT (vertical component) that were used for calculation of earthquake source parameters. Corner frequencies  $f_0$  and zero-frequency spectral density  $S_0$  are shown. Arrows show the sample windows for spectra. *Courtesy of Japan Meteorological Agency.*

### 9.2.2.1 Seismic Moment–Magnitude Relationship

Fig. 9.15 shows a “seismic moment–magnitude” dependence for the volcano-tectonic earthquakes related to central and lateral eruptions. In Fig. 9.15A, the parameters were taken from the Harvard CMT database for large 1980–96 volcano-tectonic earthquakes with magnitude  $M_s \geq 5.0$ . It is seen that the seismic moments of volcano-tectonic earthquakes related to the central eruptions of Mount St. Helens and Pinatubo volcano are larger at the same magnitudes than the seismic moments of volcano-tectonic earthquakes related to the flank eruptions of Oshima, Miyakejima, Teishi Knoll, and Grimsvøtn volcanoes. The same tendency was observed at the basaltic volcano Alaid, Kurile Islands (Fig. 9.15B) for smaller volcano-tectonic earthquakes related to its 1972 flank and 1981 central eruptions (Zobin, 1988).



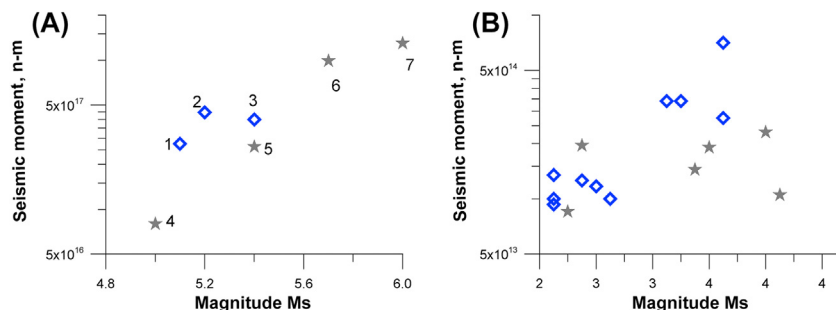


**FIGURE 9.15** Seismic moment–magnitude relations (A) for large, ( $M \geq 5$ ) volcano-tectonic earthquakes related to the central eruptions of volcanoes Pinatubo (1, 2) and Mount St. Helens (3), and the flank eruptions of Teishi Knoll (4), Grimsvötn (5), Oshima (6), and Miyake-sima (7) volcanoes and (B) for the 1972 flank and 1981 central eruptions of Alaid volcano. The data for central eruptions are marked by diamonds and for flank eruptions by stars. Data were taken from the Harvard CMT database for (A) and from Zobin, V.M., 1988. Discriminations between volcanic earthquakes associated with central and lateral eruptions. *Journal of Volcanology and Geothermal Research* 34, 307–315 for (B).

The greater amount of long-period radiation expressed in larger values of seismic moments may be attributed to the difference in the formation of the earthquake source for the two types of eruptions. Apparently, during a central eruption, the magma opens rather “smooth” fractures in a medium, which is probably already broken and possibly also hotter from the repetitive intrusion of magma along the same path. During a flank eruption, a new eruptive fracture representing a branch of the main central channel is formed every time before the eruption, which causes the generation of asperities or barriers on the rupture. These may cause a higher amount of short-period waves (Zobin, 1988).

### 9.2.2.2 Fault Area–Seismic Moment Relationship

Fig. 9.16 shows the relation between the circular fault area and seismic moments (circular dislocation Brune’s model) for volcano-tectonic earthquakes associated with the eruptions at Volcán de Colima, Etna, and Usu volcanoes (Patané et al., 1995; Zobin et al., 2002b, 2005a,b; Giampiccolo et al., 2007). This plot demonstrates the absence of a linear relation between these two source parameters. The foci are characterized by a large range of stress drops changing from 0.01 to 100 MPa. Analysis of the frequency distribution of the logarithm of stress drops of the events presented in the largest sample of the events, presented in Fig. 9.16, which was associated with the 2000 eruption of Usu volcano (Zobin et al., 2005a,b), showed a significant bimodal distribution, with stress drops lower or higher than 10 MPa (Fig. 9.17). The high-stress drop (HSD) events are more numerous (210 events) and are characterized by the mean geometrical value of  $\log \Delta\sigma = 1.59 \pm 0.24$ . The low-stress drop (LSD) 32 events are characterized by the mean geometrical value of  $\log \Delta\sigma = 0.37 \pm 0.43$ .

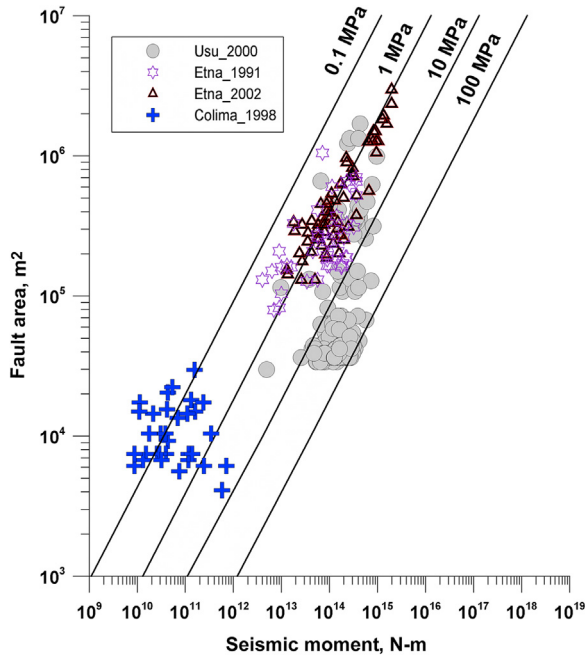


**FIGURE 9.16** Relation between seismic moment and fault area for volcano-tectonic earthquakes associated with the eruptions at Volcán de Colima (1977–78), Etna (1991, 2002), and Usu (2000) volcanoes. The lines of equal stress drop are shown. Data were taken from Patané, G., Coco, G., Corrao, M., Imposa, S., Montalto, A., 1995. Source parameters of seismic events at Mount Etna volcano, Italy, during the outburst of the 1991–1993 eruption. *Physics of the Earth and Planetary Interiors* 89, 149–162; Zobin, V.M., González Amezcu, M., Reyes-Dávila, G.A., Domínguez, T., Cerda Chacón, J. C., Chávez Álvarez, J.M., 2002b. The comparative characteristics of the 1997–1998 seismic swarms preceding the November 1998 eruption of Volcán de Colima, Mexico. *Journal of Volcanology and Geothermal Research* 117, 47–60; Zobin, V.M., Nishimura, Y., Miyamura, J., 2005a. The nature of volcanic earthquake swarm preceding the 2000 flank eruption at Usu volcano, Hokkaido, Japan. *Geophysical Journal International* 163, 265–275; Zobin, V.M., Orozco-Rojas, J., Reyes-Dávila, G.A., Navarro, C., 2005b. Seismicity of an andesitic volcano during block-lava efusión: Volcán de Colima, México, November 1998–January 1999. *Bulletin of Volcanology* 67, 679–688 and Giampiccolo, E., D'Amico, S., Patané, D., Gresta, S., 2007. Attenuation and source parameters of shallow microearthquakes at Mt. Etna Volcano, Italy. *Bulletin of the Seismological Society of America* 97, 184–197.

### 9.3 TEMPORAL VARIATIONS OF THE SOURCE SPECTRAL CHARACTERISTICS AND FOCAL MECHANISMS OF VOLCANO-TECTONIC EARTHQUAKES IN THE COURSE OF VOLCANIC ACTIVITY

#### 9.3.1 Corner Frequencies Variations

The detailed study of the body-wave spectra of volcano-tectonic earthquakes was carried out before the November 1998 andesitic lava extrusion at

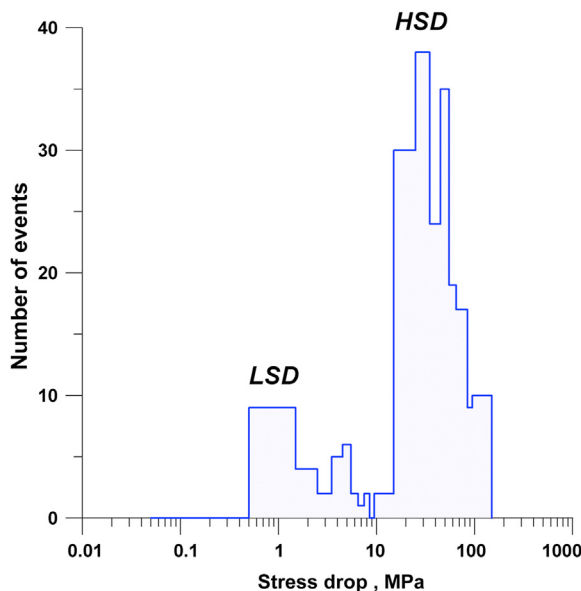


**FIGURE 9.17** Stress drop distribution for volcano-tectonic earthquakes associated with the 2000 eruption of Usu volcano. *LST*, low-stress drop events; *HST*, high-stress drop events. Data from the catalog prepared by V. Zobin.

Volcán de Colima, Mexico (Domínguez et al., 2001; Zobin et al., 2002a,b,c). Five earthquake swarms occurred during 1 year of seismic activity, from November 1997 to November 1998 (see Section 6.3.1). For *P*-wave spectra, two groups of events were observed: low-frequency (LF) events with values of  $f_0$  from 2 to 7 Hz and high-frequency (HF) events with values of  $f_0$  from 8 to 15 Hz (Fig. 9.18B). These two groups were separated at the significance level of 99%. They were observed simultaneously during all five swarms (Fig. 9.18A). The same two groups of LF and HF events were observed also for S-wave spectra during the June 1988–July 1988 swarm 4 when a three-component seismic station was in operation.

The relative abundances of LF and HF events changed from swarm to swarm (Fig. 9.18A). HF events dominated swarms 1 and 2, whereas LF events were twice as abundant as HF events in swarm 4. The numbers of LF and HF events were practically equal for swarm 5, as they were for swarm 3. Zobin et al. (2002b) noted that a sharp decrease in LF events was observed 2 days before the Colima eruption.

The two groups of HF and LF events were observed also for volcano-tectonic earthquakes related to the 1975–76 Tolbachik (see Section 5.3.1) and the 1991–93 Etna (see Section 5.2.1) eruptions (Fig. 9.19). For New Tolbachik volcanoes, the HF events ( $f_0 > 2.5$  Hz) were observed only before the opening of the fissure when the first cone was formed, 6 days before the

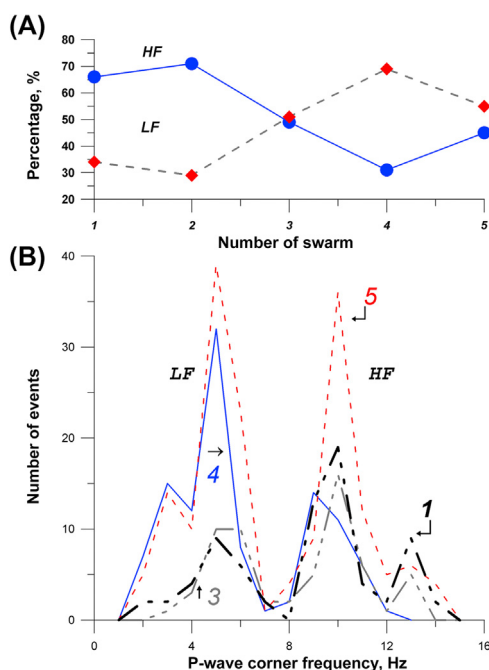


**FIGURE 9.18** Variations in the abundance of LF and HF events from swarm to swarm (A) and distribution of events with different  $P$ -wave corner frequencies during swarms 1, 3, 4, and 5 (B) Arrows show the eruptive events. LF, low frequency; HF, high frequency. From Zobin, V.M., González Amezcua, M., Reyes-Dávila, G.A., Domínguez, T., Cerda Chacón, J.C., Chávez Álvarez, J.M., 2002b. The comparative characteristics of the 1997–1998 seismic swarms preceding the November 1998 eruption of Volcán de Colima, Mexico. *Journal of Volcanology and Geothermal Research* 117, 47–60.

beginning of the eruption. During the two swarms that occurred before the eruptions of the third and eighth cones, only the LF events were observed.

For the December 1991 Etna events, the two groups of LF and HF ( $f_0 > 2.5$  Hz) events occurred simultaneously during the sharp increase in seismic activity at the initial stage of swarm activity on December 14 (Fig. 9.19B). With the end of peak activity and the opening of the eruptive fissure, LF events disappeared for 2 hours. Then both groups appeared again up to the beginning of the lava flow on December 15.

Zobin et al. (2002a,b,c) suggested that a moving magma dyke could produce HF seismic events at distances less than 70 m from the dyke tip along existing fractures and trigger the fracturing of local faults at distances greater than 70–100 m from the dyke tip as LF events. Therefore the occurrence of HF events may indicate the moving of magma from depths, and the LF events reflect the process of passageway construction.

From this point of view, a sharp decrease in LF events 2 days before the Colima 1998 lava eruption indicates a magma uplift before the 1998 lava

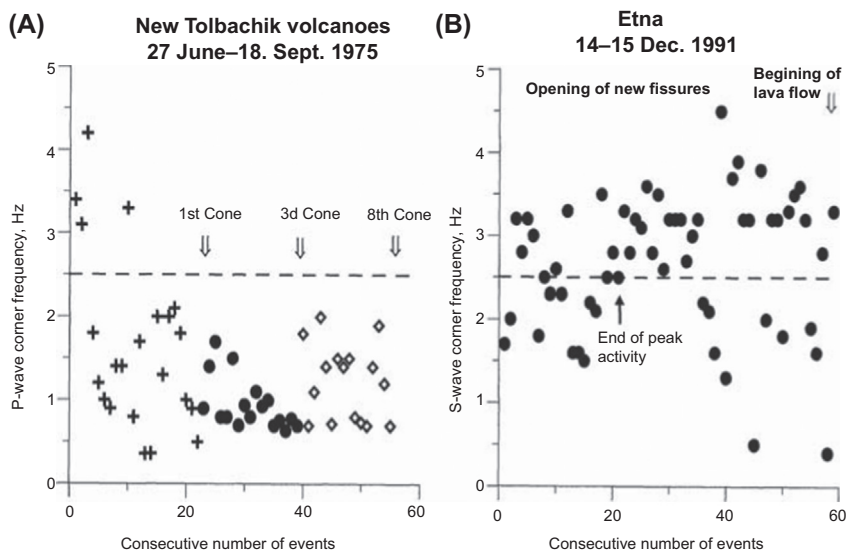


**FIGURE 9.19** Temporal variations of P-wave corner frequencies for volcano-tectonic earthquakes associated with the 1975 fissure eruption of New Tolbachik volcanoes, Kamchatka (A) and of S-wave corner frequencies for volcano-tectonic earthquakes associated with the 1991 Etna flank eruption (B) Arrows show the eruptive events. Data from Zobin, V.M., Chirkova, V.N., 1984. *Catalog of source parameters of earthquakes based on P-wave spectra (Klyuchi station records)*. In: Fedotov, S.A. (Ed.), *Large Tolbachik Fissure Eruption*, pp. 607–609. Nauka, Moscow. (in Russian) and Patané, G., Coco, G., Corrao, M., Imposi, S., Montalto, A., 1995. *Source parameters of seismic events at Mount Etna volcano, Italy, during the outburst of the 1991–1993 eruption. Physics of the Earth and Planetary Interiors* 89, 149–162.

eruption along the prepared passageway. For the New Tolbachik volcanoes, magma may be considered as arriving close to the surface on July 30, 1975, when the HF events disappeared; during the following 6 days, a passageway was prepared along the new fissure, then the eruption at the first cone began. During the forming of the third and eighth cones, magma did not rise from depth, rather it moved laterally from the site of the first cone and from Plosky Tolbachik (see Section 5.3.1). Therefore only LF events, that ruptured the fissure for new cones, were recorded.

### 9.3.2 Stress Drop Variations

Fig. 9.20 shows that the 2000 Usu earthquakes (see Section 7.2.1) were separated into two groups as characterized by HSD and LSD. The temporal variations of the volcano-tectonic earthquakes with different stress drops were studied corresponding to the four stages in crustal deformation. These were recorded simultaneously with the seismic activity by the GPS stations situated at a distance of 2–3 km from the central crater. During the first stage

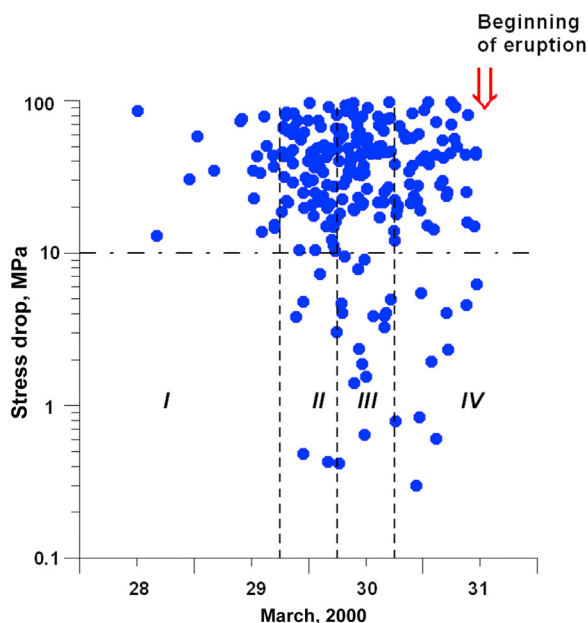


**FIGURE 9.20** Temporal variations of the 2000 Usu volcano-tectonic earthquakes with different stress drops. I–IV, the stages of deformation, according to Takahashi et al. (2002). Stage I: March 27, 20:00 h to March 29, 18:00 h, no deformations. Stage II: March 29, 18 h to March 30, 06 h, beginning of slow deformation. Stage III: March 30, 06:00 h to March 30, 18:00 h, sharp increase in deformation rate. Stage IV: March 30, 18:00 h to March 31, 13:00 h, slow deformation near the volcano and fast deformation near the site of forthcoming eruption (KON); the eruption began at the end of the stage. *Double arrows* show the beginning of the eruption. The *single arrows* show the day duration. Data from the catalog prepared by V. Zobin.

(without any deformation), only HSD events occurred. The LSD events began to occur about 30–40 hours after the beginning of the deformation of the volcanic edifice. Then the two groups of volcano-tectonic earthquakes were recorded during the third and fourth stages. All the time, the HSD events were predominant in number (Zobin et al., 2005a).

Fig. 9.21A shows the variations of LSD (<1 bar) and HSD volcano-tectonic earthquakes preceding the 1975 Tolbachik eruption. It is seen that both LSD and HSD events were recorded before the occurrence of two  $M_w$  4.9 earthquakes on July 2. Then, HSD events disappeared and only LSD events were observed till the beginning of the eruption (Zobin and Gorelchik, 1982).

Zobin et al. (2005a) attributed the HSD earthquakes as the probable product of magmatic heat fracturing of host rocks. The relatively long LSD ruptures may be attributed to the slips along the preexisting and forming due to the occurrence of large earthquake fractures during magma movement.

Therefore the knowledge of temporal variations of the spectral source parameters may allow the description of the main events during the eruptive process.

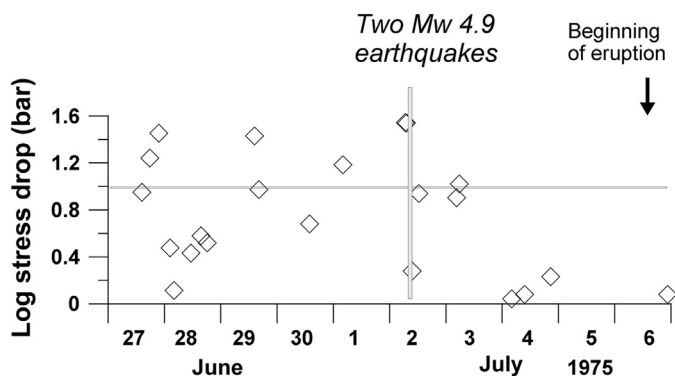


**FIGURE 9.21** Variations of the LSD ( $<1$  bar) and HSD events for volcano-tectonic earthquakes preceding the 1975 Tolbachik eruption. The time of occurrence of two July 2  $M_w$  4.9 earthquakes is shown by the vertical line. The horizontal line shows the separation of the LSD and HSD events at the level of 1 bar. *LST*, low-stress drop events; *HST*, high-stress drop events. Data from Zobin, V.M., Chirkova, V.N., 1984. *Catalog of source parameters of earthquakes based on P-wave spectra (Klyuchi station records)*. In: Fedotov, S.A. (Ed.), *Large Tolbachik Fissure Eruption*, pp. 607–609. Nauka, Moscow. (in Russian).

### 9.3.3 Stress Field Rotations

The stress field rotations obtained from focal mechanism solutions were recorded during the swarms of volcano-tectonic earthquakes associated with the 1975 Tolbachik fissure eruption in Kamchatka (Zobin, 1979b; Zobin and Gorelchik, 1982), the 1992 Crater Peak eruption in Alaska (Roman et al., 2004), and the 2004–05 Mount St. Helens eruption (Lehto et al., 2010).

During the 1975 Tolbachik swarm, this rotation was observed just after two  $M_w$  4.9 earthquakes had occurred on July 2 (see Fig. 4.20 from Chapter 4) simultaneously with the disappearance of HSD earthquakes (see Fig. 9.21). It continued up to the beginning of the eruption on July 6.

The 2004–08 eruption at Mount St. Helens was preceded by a swarm of shallow volcano-tectonic earthquakes that began on September 23, 2004 (see Section 7.1.1). The focal mechanism solutions were calculated for shallow volcano-tectonic earthquakes recorded during a background period (January 1999–July 2004) and during the first day of the eruption (early vent-clearing phase, September 23, 2004, to September 29, 2004). Fig. 9.22 shows the normal and SS faulting during the background period and on September 23; SS and reverse faulting on September 24; and a mixture of SS, reverse, and normal faulting on September 25–September 29. The orientation of  $\sigma_1$  beneath the volcano, as estimated from stress tensor inversions, was found to be subhorizontal for all periods and oriented NE–SW during the background period, NW–SE on September 24, and NE–SW on September 25–September 29. It was suggested that the ephemeral  $\sim 90$ -degree change in  $\sigma_1$  orientation was due to intrusion and inflation of an NE–SW-oriented dyke in the shallow crust before the eruption onset (Fig. 9.23).

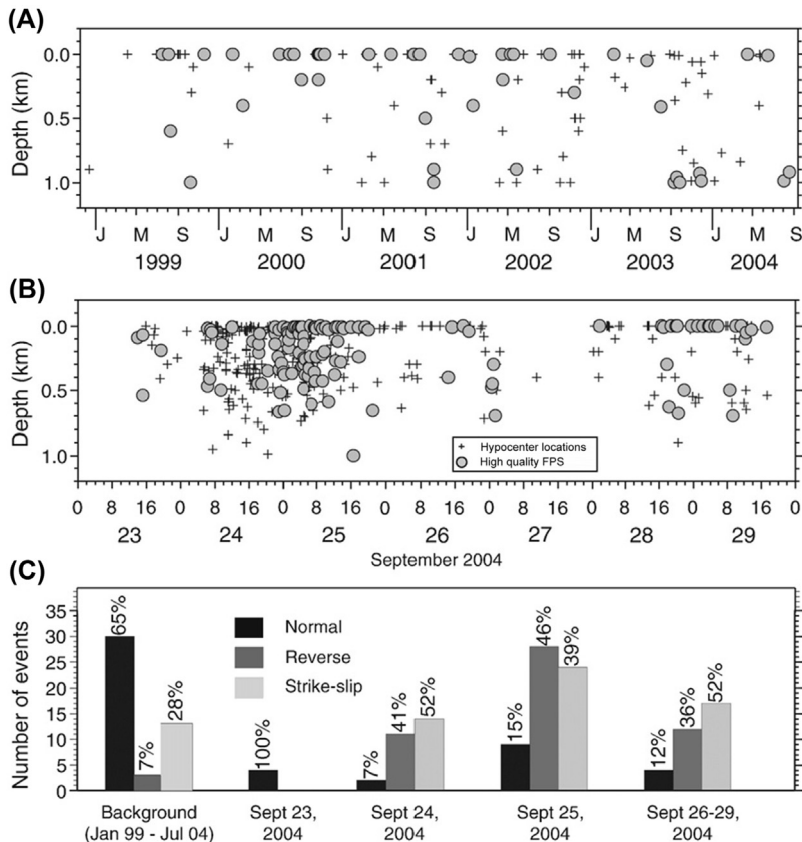


**FIGURE 9.22** Variations in the fault nature of volcano-tectonic earthquakes during the background and the initial vent-clearing phase of the 2004 Mount St. Helens eruption. (A) Time–depth plot of hypocenters for all well-located volcano-tectonic earthquakes (*black crosses*) and for well-located volcano-tectonic earthquakes with well-constrained focal mechanism solutions (*gray circles*) for the background period. (B) Time–depth plot of hypocenters for all well-located volcano-tectonic earthquakes and for those with well-constrained focal mechanism solutions for the vent-clearing phase, September 23, 2004 through September 29, 2004. (C) Histograms of faulting types as defined in the text. Note the sharp decrease in normal faulting on September 24 and its reappearance on September 25. Percentages of faulting type for each time period are also shown (e.g., four September 23 events show 100% normal faulting). The background period covers a time span of ~5 years, while the vent-clearing period covers a time span of 7 days. All depths are relative to a datum of 2.2 km a.s.l. From Lehto, H.L., Roman, D. C., Moran, S.C., 2010. Temporal changes in stress preceding the 2004–2008 eruption of Mount St. Helens, Washington. *Journal of Volcanology and Geothermal Research* 198, 129–142.

## 9.4 SEISMOTECTONIC DEFORMATIONS IN THE VOLCANIC REGION

Knowledge of the focal mechanisms and scalar seismic moments of volcano-tectonic earthquakes, associated with an eruption, allows the



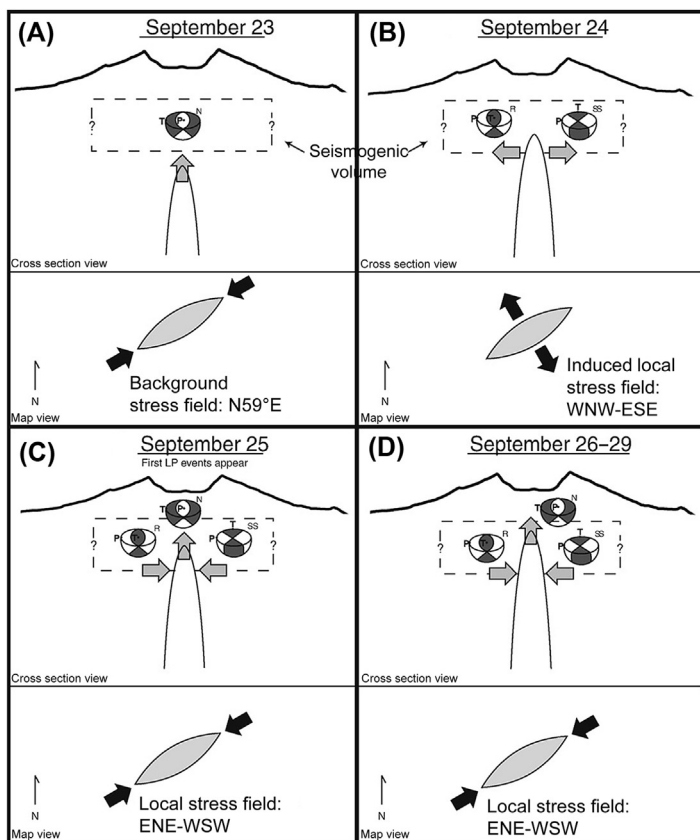


**FIGURE 9.23** A proposed model of intrusion and inflation of a magma-filled dyke into the shallow subsurface during the initial phase of the 2004 Mount St. Helens eruption shown in both cross-section and map view for: (A) September 23, 2004, (B) September 24, 2004, (C) September 25, 2004, and (D) September 26, 2004 to September 29, 2004. Gray arrows represent either upward movement (*vertical arrows*) or inflation/deflation (*horizontal arrows*) of the dyke, while black arrows show the orientation of maximum compression. N, R, and SS stand for normal faulting, reverse faulting, and strike-slip faulting events, respectively. Dots labeled *P* and *T* represent the positions of the *P*- and *T*-axes, respectively. Dashed box represents the approximate location of the seismogenic volume hosting the volcano-tectonic events. Lower hemisphere projections of focal mechanisms shown are schematic and do not necessarily reflect the exact hypocenter locations. From Lehto, H.L., Roman, D.C., Moran, S.C., 2010. Temporal changes in stress preceding the 2004–2008 eruption of Mount St. Helens, Washington. *Journal of Volcanology and Geothermal Research* 198, 129–142.

determination of seismotectonic deformations (STD) within the volcanic edifice and surroundings. It is produced by the process of magma deep migration before, during, and after the eruption. STD  $\varepsilon_{ik}$  of the volume

$V$  that contained the earthquake foci can be estimated (Riznichenko, 1977) using the following equation:

$$\varepsilon_{ik} = \frac{\sum Mo(\theta_{ik})}{2\mu V} \quad (9.4)$$

where  $\mu$  is accepted to be  $3.2 \times 10^{11}$  dyne/cm<sup>2</sup>;  $V$  is equal to 30 km<sup>3</sup>;  $\theta_{ik}$  is the unit tensor of seismic moment  $Mo$ ; and  $\Sigma$  is the symbol of sum.

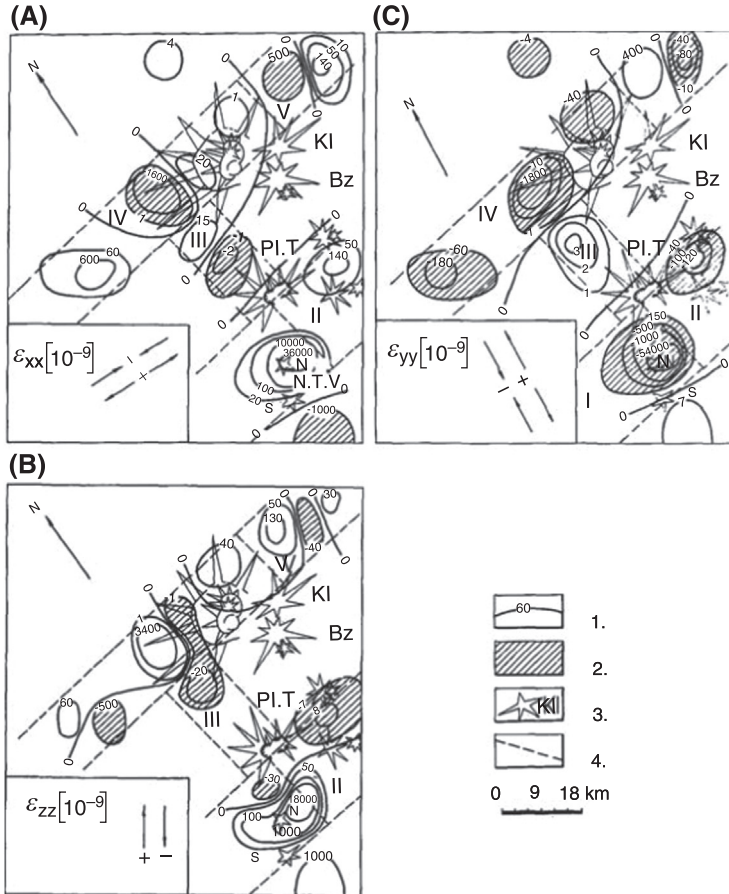


**FIGURE 9.24** Distribution of extensional deformations in the region of the 1975–76 Tolbachik fissure eruption, Kamchatka in latitudinal (east–west) (A), meridional (north–south) (B) and vertical (C) directions. *KI*, Klyuchevskoy; *Bz*, Bezymianny; *PIT*, Plosky Tolbachik; *NTV*, New Tolbachik volcanoes *N* and *S* indicated the north and south groups of new cones of New Tolbachik volcanoes. (1) The isolines of deformations, (2) the regions of negative deformations, (3) the volcanoes, and (4) a boundary of zones I–V. From Zobin, V.M., 1990. Seismotectonic crustal deformations during the Large Tolbachik fissure eruption. *Journal of Volcanology and Geothermal Research* 43, 271–278.

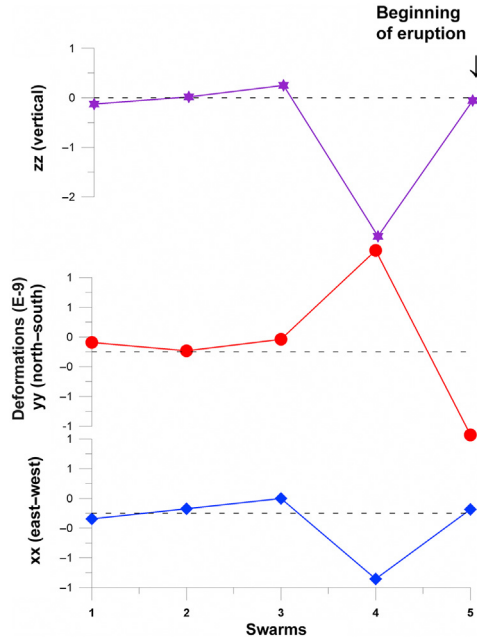


**FIGURE 9.25** Variations of principal components.  $\varepsilon_{xx}$ ,  $\varepsilon_{yy}$ , and  $\varepsilon_{zz}$  of seismotectonic deformations (STD) from swarm to swarm 1 year before the November 1998 lava eruption at Volcán de Colima 1–5 denotes the numbers of swarms according to their temporal appearance. Zero line is shown by the dashed line. From Zobin, V.M., González-Amezcu, M., Reyes-Dávila, G.A., 2002a. Seismotectonic deformations of volcanic edifice prior to the 1998 eruption of Volcán de Colima, México. *Bulletin of Volcanology* 64, 349–355.

The components  $\theta_{xx}$ ,  $\theta_{yy}$ , and  $\theta_{zz}$  of the unit tensor of the seismic moment are estimated using focal mechanism parameters through formulae

$$\theta_{xx} = \sin^2 \sigma_T \sin^2 \alpha_T - \sin^2 \sigma_P \sin^2 \alpha_P \quad (9.5)$$

$$\theta_{yy} = \sin^2 \sigma_T \cos^2 \alpha_T - \sin^2 \sigma_P \cos^2 \alpha_P \quad (9.6)$$

$$\theta_{zz} = \cos^2 \sigma_T - \cos^2 \alpha_P \quad (9.7)$$

where  $\sigma_P$  and  $\sigma_T$  are the angles of  $P$ - and  $T$ -axes with the vertical;  $\alpha_P$  and  $\alpha_T$  are the azimuthal angles of the  $P$ - and  $T$ -axes. The  $X$ -axis is oriented to the east, the  $Y$ -axis to the north, and the  $Z$ -axis upward. The STD was calculated for two volcanic zones: for the 1975–78 volcano-tectonic activity in the region of Tolbachik fissure eruption, Kamchatka (Zobin, 1990) and for the 1997–98 volcanic pre-lava unrest at Volcán de Colima, Mexico (Zobin et al., 2002a). Figs. 9.24 and 9.25 illustrate the results.

The 1975–78 seismovolcanic activity in the region of Tolbachik fissure eruption was characterized by strong 1975–76 volcanic activity along the new fissure and by the following posteruptive seismic activity within the surroundings

(see Section 5.3.1). The magnitude 4–5 seismic events were recorded at the site of the cones just before the beginning of the eruption and at a distance of 3–50 km from the active cones just after the ceasing of the eruption. Fig. 9.24 shows the spatial distribution of the three components  $\varepsilon_{xx}$ ,  $\varepsilon_{yy}$ , and  $\varepsilon_{zz}$  of extensional deformations that were calculated for all recorded earthquakes of magnitude  $m_b > 3.7$  that occurred within the upper crust layer from 0 to 20 km. In the eruptive zone I, the largest compressional deformations dominated with a value of  $-5.4 \times 10^{-5}$ ; while the largest extensional deformations were observed in the latitudinal direction with the maximum value of  $3.6 \times 10^{-5}$ . The vertical deformations reached to  $1.6 \times 10^{-5}$ . In the periphery, the maximum deformations were observed in zone IV situated 40 km to the north of the eruptive sites, where the compression was dominant reaching a value of  $-1.8 \times 10^{-6}$  while the vertical deformation was of  $3.4 \times 10^{-6}$ .

Generally, the comparison of seismic activity development (see Section 5.3.1) and the scheme of Fig. 9.24 allow the reconstruction of the process of total crustal deformation within the seismic volume in the region of the fissure eruption. The seismic activity started in zone I, the eruption zone. This led to considerable bulging of the Earth's surface producing extensional deformations in latitudinal and compressional in meridional directions. Later, after the termination of the eruption the seismic activity spread radially from Plosky Tolbachik (zones II and III in Fig. 9.24). Zone II extends along the Upper Pleistocene–Holocene extrusive domes to the east of Plosky Tolbachik. Zone III extends along the western border of the distribution of the Upper Pleistocene–Holocene cinder and lava cones. Zones II and III were notable for subsidence of different amplitudes that can be caused by the discharge of hydraulically connected magma chambers (Fedotov, 1984b). If this assumption is correct, then the outlines of the subsidence zone in Fig. 9.24C can serve as the boundary of the zone of the near-surface magma chambers that feed the Klyuchevskaya group of volcanoes. Collapses of the near-surface rocks along the margin of the magma chambers could obviously produce compensational bulging in zones IV and V that resulted in the upsurge of seismic activity in 1977–78 (Zobin, 1990).

The 1997–98 pre-lava eruption unrest of Volcán de Colima, Mexico (see Section 6.3.1) was more local and produced lesser deformations, mainly within the volcanic edifice from sea level to the crater at the altitude of 3860 m (Zobin et al., 2002a). Fig. 9.25 shows the temporal variations of three components of extensional deformations 1 year before the eruption. The calculated deformations vary from  $1.3 \times 10^{-11}$  to  $2.7 \times 10^{-9}$ . The maximum deformation was observed in June 1998–July 1998 (swarm 4), and the minimum deformation was observed in March 1998 (swarm 2). The process of deformation was rather complicated. The first earthquake swarm of November 1997–December 1997 resulted in an N–S horizontal elongation that is twice greater than the E–W horizontal shortening. The volume was also subject to vertical shortening. In March

1998 and May 1998 (swarms 2 and 3), uniform horizontal N–S and E–W elongations, accompanied by vertical elongation of the volume were observed. In June 1998–July 1998 (swarm 4), the situation of November 1997–December 1997 (swarm 1) was repeated with the N–S horizontal elongation greater than the EW horizontal shortening of the volume and with an intensive vertical shortening. During the last swarm of October 1998–November 1998, the slight E–W elongation of the seismic volume was accompanied by strong N–S shortening and a very slight vertical shortening.

Quantification of the deformation process within the volume for seismic foci recorded before the November 1998 eruption of Volcán de Colima enables interpretation of the process, thus giving volcanological meaning to the deformation terms. The seismic volume can be considered as the volume of the volcanic edifice. The process of horizontal elongation is interpreted as the process of opening a magma passageway; vertical shortening in the seismic volume is interpreted to reflect magma ascent. The N–S component of deformation coincides with the orientation of the Colima rift and the E–W component of deformation can be associated with the adjusting NE–SW Tamazula fault system (Zobin et al., 2002a).

Based on these assumptions, magma began to rise in November–December 1997 forming a passageway along the Tamazula fault system. In March 1998 and May 1998, the vertical movement of magma practically stopped, and magma began to fill existing fractures in both the Colima rift and the Tamazula fault system. In June 1998–July 1998, a new strong phase of magma ascent began and the passageway was formed along the Tamazula fault system. During the last swarm in October 1998–November 1998 the final passageway was formed, with an N–S orientation within the Colima rift system, and the final ascent of magma to the surface was completed (Zobin et al., 2002a).

The study of STD gives us the possibility for a quantitative comparison of two eruptions. The larger magnitude of events recorded during the seismic activity preceding the Tolbachik eruption yielded larger values of STD that reached  $10^{-5}$ . This difference in four orders of magnitude with the Colima deformations agrees well with the difference in the magmatic volume of erupted products  $2.2 \text{ km}^3$  for the Tolbachik eruption versus  $0.04 \text{ km}^3$  for the 1998–99 Colima eruption (Fedotov, 1984a; Zobin et al., 2002c).

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## RELEVANT WEBSITE

Harvard (CMT). <http://www.globalcmt.org/CMTsearch.html>.

# Significant Volcano-Tectonic Earthquakes and Their Role in Volcanic Processes

Omori (1912) wrote that a volcanic earthquake could never be equal to a large tectonic shock. He considered that the zones of active volcanoes “may be regarded as a sort of safety valves to, or lines of least resistance in, the Earth’s crust, preventing an accumulation in their localities of underground stress sufficient to result in a gigantic earthquake. Hence it is that extremely violent or very extensive earthquakes never originate from under volcanoes.”

Long-term observations showed that volcanic earthquakes are characterized by significantly lower  $M_{\max}$  and a lower frequency of large earthquake appearance than tectonic earthquakes. Most volcanic earthquakes are small; their magnitudes do not exceed 2 or 3. At the same time, significant (moderate-size and large) volcanic earthquakes may occur and produce a local effect of high intensity (Abe, 1979; Yokoyama, 2001; Zobin, 2001).

Machado et al. (1962) described the macroseismic effects that were produced by the May 13, 1958, to May 14, 1958, volcano-tectonic earthquake sequence at Fayal Island during the flank eruption of Fayal volcano. About 450 earthquakes were felt on Fayal; many of them had a Modified Mercalli (MM) scale intensity of X (Fig. 10.1A). Of course, these effects were very local, covering an area of about 2–3 km<sup>2</sup>, and they were strongly multiplied by a surface layer of soft volcanic rocks on the flanks of the volcano. Fig. 10.1B shows the macroseismic effect produced by the  $M_L$  3.6 earthquake occurring within the edifice of Vesuvius volcano on October 9, 1999 (Cubellis et al., 2007). It was felt over a wide area within about 25 km from the crater and caused fear and anxiety among the people. A macroseismic study of this event was carried out from an analysis of questionnaires sent out to all secondary schools in each municipality of the Vesuvian and Neapolitan areas and surrounding towns. The contours of felt index areas illustrate that maximum intensities, estimated as VI MCS



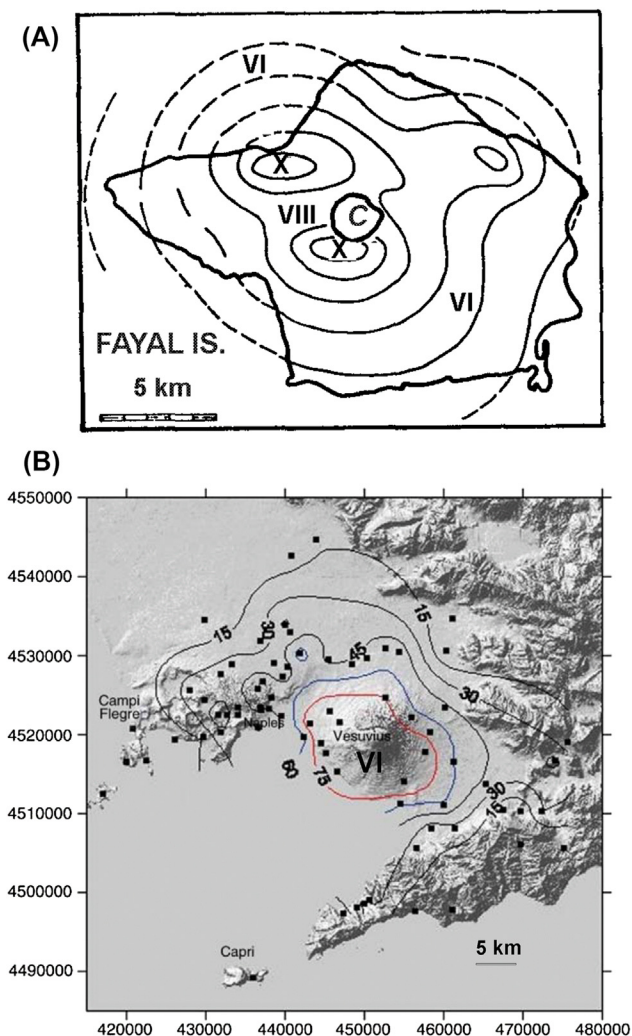


**FIGURE 10.1** Macroseismic effect of the volcano-tectonic earthquakes observed (A) at Fayal volcano during the May 13, 1958 to May 14, 1958 seismic crisis associated with the flank eruption of the volcano and (B) at Vesuvius volcano during the largest event of the August 1999–October 1999 seismic sequence. In (A), the numbers VI to X are the Modified Mercalli scale intensities; C indicates the caldera (From Machado, F., Parsons, W.H., Richards, A.F., Mulford, J.W., 1962. Capelinhos eruption of Fayal volcano, Azores, 1957–1958. *Journal of Geophysical Research* 67, 3519–3529.). In (B), the Mercalli–Cancani–Sieberg scale (MCS) intensity VI was estimated as  $I_0$  for the magnitude  $M_L$  3.6 earthquake occurring on October 9, 1999. Felt index map is shown. The felt index is the percentage response to the question, Did you feel the earthquake? The area of maximum value (75%) covers the whole Vesuvian area. The felt index weakens strongly in NE and SE directions (From Cubellis, E., Luongo, G., Marturano, A., 2007. Seismic hazard assessment at Mt. Vesuvius: maximum expected magnitude. *Journal of Volcanology and Geothermal Research* 162, 139–148).

(Mercalli—Cancani—Sieberg scale, VI MCS = V MM; Tiedemann, 1992), correspond to the volcanic edifice from which they decrease radially.

Blong (1984) wrote about at least 29 deaths associated with the secondary effects (landslides, building collapse, etc.) of volcanogenic earthquakes during the eruptions of Sakurajima, Kyushu (1914) and Api Siau, Indonesia (1976). Therefore a seismic hazard of volcano-tectonic earthquakes for the settlements on the slope of the volcano may be a reality.

## 10.1 SELECTION OF SIGNIFICANT VOLCANO-TECTONIC EARTHQUAKES

Table 10.1 lists the significant ( $M \geq 4.5$ ) earthquakes, which occurred near volcanoes from 1910 to 2005. They are very rare, rarer than magnitude 7.5–8.5 tectonic earthquakes. Only three magnitude  $M_s$  7.0–7.1 volcanic earthquakes were recorded during the century (Abe, 1979, 1992), while the annual average number of  $M_w > 7.0$  tectonic events is about 15 (Lay and Wallace, 1995).

The catalog of Table 10.1 contains 32 events. Of the 32 events, 3 events are with magnitude  $M_w \geq 7.0$ , 4 events are with magnitude from 6.0 to 6.9, and 17 events are with magnitude from 5.0 to 5.9. This catalog was prepared for significant seismic events associated with 32 eruptions of 26 volcanoes (Fig. 10.2). The earthquakes were associated with eruptions of different types, and they occurred at different stages of the eruptions. It is unlikely that it is the complete catalog for the century but it includes all available published data. Unfortunately, the large amount of information about the seismic events of the large eruptions before the 1960s was out of publication.

Only the largest earthquakes for each eruption are listed in Table 10.1. These selected earthquakes had different magnitudes, such as  $M_s$ ,  $mb$ ,  $MLH$ ,  $M_w$ , and  $K_s$ . It was necessary to prepare the homogeneous catalog according to a single magnitude. The magnitude  $mb$  was not representative, since it became saturated at the magnitude  $mb = 5.8$  (Fig. 10.3). From two other magnitudes,  $M_s$  and  $M_w$ , the magnitude  $M_w$  was used as a basic. The moment magnitudes  $M_w$  estimated by the Harvard Centroid-Moment Tensor (CMT) group and by the authors (Zobin, 1979; Zobin, 1988; Zobin and Levina, 1998) and the International Seismological Centre (ISC) magnitudes  $M_s$  for 10 events were used for the construction of the relationship between  $M_w$  and  $M_s$  for volcanic earthquakes (Fig. 10.4). It was obtained using Eq. (10.1)

$$M_w = 0.86 M_s + 1.00 \quad (10.1)$$

The above equation was used for the classification of earthquakes with an unknown seismic moment. The final magnitude limit for the catalog is  $M_w \geq 4.5$ .

**TABLE 10.1** List of Significant (Moderate-Size and Large) Volcanic Earthquakes, 1910–2005.

| No. | yyyymmdd | hhmm | Latitude (degree)   | Longitude (degree) | Depth (km) | <i>mb</i>         | <i>M<sub>s</sub></i> | <i>M<sub>w</sub></i> | Volcano      |
|-----|----------|------|---------------------|--------------------|------------|-------------------|----------------------|----------------------|--------------|
| 1   | 19100624 | 0749 |                     |                    |            |                   | 5.1 <sup>1</sup>     | 5.4*                 | Usu          |
| 2   | 19120608 | 0736 |                     |                    |            | 6.8 <sup>2</sup>  | 7.0 <sup>2</sup>     | 7.0*                 | Katmai       |
| 3   | 19140112 | 1028 |                     |                    |            |                   | 7.0 <sup>1</sup>     | 7.0*                 | Sakurajima   |
| 4   | 19430215 | 0139 |                     |                    |            |                   | 4.5 <sup>3</sup>     | 4.9*                 | Parícutin    |
| 5   | 19560330 | 0611 |                     |                    |            |                   | 5.0 <sup>4</sup>     | 5.3*                 | Bezymianny   |
| 6   | 19620826 | 0648 |                     |                    |            |                   |                      | 6.1 <sup>5</sup>     | Miyakejima   |
| 7   | 19641111 | 1906 |                     |                    |            | 5.5 <sup>7</sup>  |                      | 5.2 <sup>6</sup>     | Sheveluch    |
| 8   | 19680613 | 0733 |                     |                    |            | 5.4 <sup>7</sup>  |                      | 5.4 <sup>6</sup>     | Fernandina   |
| 9   | 19680717 | 0623 |                     |                    |            | 5.1 <sup>7</sup>  | 4.8**                | 5.1*                 | Arenal       |
| 10  | 19700918 | 0206 |                     |                    |            | 5.1 <sup>7</sup>  | 5.3 <sup>12</sup>    | 5.7 <sup>5</sup>     | Beerenberg   |
| 11  | 19730711 | 2247 |                     |                    |            | 5.3 <sup>7</sup>  | 5.0 <sup>12</sup>    | 5.9 <sup>8</sup>     | Curacoa Reef |
| 12  | 19730714 | 0115 |                     |                    |            | 4.2 <sup>7</sup>  |                      | 4.6 <sup>5</sup>     | Tiatia       |
| 13  | 19740129 | 1334 | <sup>9</sup> 37.77  | 15.08              | 15         | 4.5 <sup>13</sup> | 4.3 <sup>9</sup>     | 4.7*                 | Etna         |
| 14  | 19750702 | 0734 | <sup>15</sup> 55.41 | 160.15             | 15         | 4.7 <sup>7</sup>  |                      | 4.9 <sup>6</sup>     | Tolbachik    |
| 15  | 19770106 | 1833 |                     |                    |            | 5.2 <sup>7</sup>  |                      | 5.1 <sup>10</sup>    | Nyiragongo   |
| 16  | 19771004 | 1339 |                     |                    |            |                   | 4.3 <sup>1</sup>     | 4.7*                 | Usu          |
| 17  | 19800518 | 1532 | <sup>16</sup> 46.21 | −122.19            | 3          | 4.6 <sup>7</sup>  | 5.2 <sup>7</sup>     | 5.4 <sup>6</sup>     | St. Helens   |
| 18  | 19830719 | 1044 |                     |                    |            | 5.2 <sup>7</sup>  | 4.5 <sup>12</sup>    | 4.9*                 | Una Una      |

|    |          |      |                     |        |    |                  |                   |                   |               |
|----|----------|------|---------------------|--------|----|------------------|-------------------|-------------------|---------------|
| 19 | 19831003 | 1333 | <sup>17</sup> 34.00 | 139.52 | 15 | 5.8 <sup>7</sup> | 6.2 <sup>7</sup>  | 6.0 <sup>10</sup> | Miyakejima    |
| 20 | 19850107 | 2153 | <sup>7</sup> 71.13  | −7.52  | 10 | 4.9 <sup>7</sup> | 4.3 <sup>7</sup>  | 5.0 <sup>10</sup> | Beerenberg    |
| 21 | 19861122 | 0041 | <sup>17</sup> 34.53 | 139.36 | 15 | 5.8 <sup>7</sup> | 5.9 <sup>7</sup>  | 5.9 <sup>10</sup> | Oshima        |
| 22 | 19890224 | 1236 |                     |        |    | 5.2 <sup>7</sup> | 4.8 <sup>7</sup>  | 5.2 <sup>10</sup> | Lonquimay     |
| 23 | 19890709 | 0209 | <sup>17</sup> 34.99 | 139.11 | 3  | 5.0 <sup>7</sup> | 5.2 <sup>7</sup>  | 5.2 <sup>10</sup> | Teisi Knoll   |
| 24 | 19910615 | 1115 | <sup>12</sup> 15.12 | 120.35 | 10 | 5.7 <sup>7</sup> | 5.4 <sup>7</sup>  | 5.6 <sup>10</sup> | Pinatubo      |
| 25 | 19910815 | 1216 |                     |        |    | 5.3 <sup>7</sup> | 5.6 <sup>7</sup>  | 5.8*              | Hudson        |
| 26 | 19911215 | 2000 | <sup>18</sup> 37.74 | 14.97  | 2  | 4.3 <sup>7</sup> | 4.3 <sup>13</sup> | 4.7*              | Etna          |
| 27 | 19960101 | 0957 | <sup>14</sup> 53.90 | 159.43 | 0  | 5.8 <sup>7</sup> | 6.7 <sup>7</sup>  | 7.1 <sup>14</sup> | Akademia Nauk |
| 28 | 19960929 | 1048 | <sup>19</sup> 64.68 | −17.49 | 4  | 5.2 <sup>7</sup> | 5.4 <sup>7</sup>  | 5.6 <sup>10</sup> | Grimsvötn     |
| 29 | 20000331 | 1812 | <sup>17</sup> 42.50 | 140.83 | 7  | 4.9 <sup>7</sup> | 4.6 <sup>7</sup>  | 4.9 <sup>20</sup> | Usu           |
| 30 | 20000701 | 0701 | <sup>17</sup> 34.21 | 139.22 | 15 | 5.9 <sup>7</sup> | 6.1 <sup>7</sup>  | 6.2 <sup>20</sup> | Miyakejima    |
| 31 | 20000730 | 1225 | <sup>17</sup> 33.96 | 139.40 | 18 | 5.9 <sup>7</sup> | 6.5 <sup>7</sup>  | 6.4 <sup>20</sup> | Miyakejima    |
| 32 | 20050924 | 1924 | <sup>7</sup> 12.39  | 40.57  | 13 | 5.0 <sup>7</sup> | 5.1 <sup>7</sup>  | 5.5 <sup>10</sup> | Dabbahu       |

*Notes:* The superscripts denote the following references: 1, [Abe, 1979](#); 2, [Abe, 1992](#); 3, Yokoyama and de la Cruz-Reyna, 1990; 4, [Okada, 1983](#); 5, Zobin, 1979a; 6, [Zobin, 1988](#); 7, ISC; 8, Gibowicz et al., 1974; 9, Bottari et al., 1975; 10, Harvard CMT; 11, Wano and Okada, 1980; 12, NEIS; 13, MOS; 14, [Zobin and Levina, 1998](#); 15, Zobin, 1979b; 16, University of Washington, Data Base; 17, JMA; 18, Ferrucci and Patane, 1993; 19, Icelandic Meteorological Office; 20, NIED, Tsukuba. (\*)  $M_w$  was calculated from magnitude  $M_s$  using the relation  $M_w = 0.86 M_s + 1.00$ ; (\*\*)  $M_s$  was calculated from the magnitude  $mb$  using the relation  $M_s = 2.5mb - 8.0$  ([Zobin, 1999](#)); () The first earthquake of magnitude  $M_s$  4.3 Usu earthquakes from a sequence of 16 events of the same magnitude which occurred from Oct. 4, 1977 to Aug. 22, 1978.

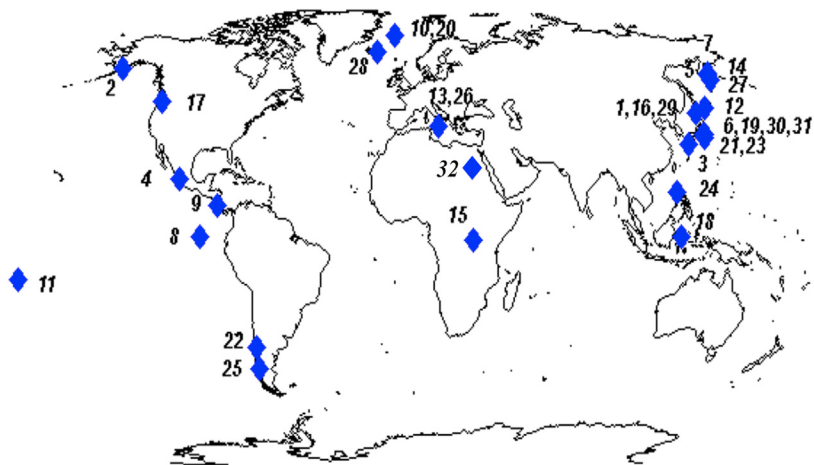


**FIGURE 10.2** The map of volcanoes under study. The volcanoes are shown by *diamonds* with numbers according to [Table 10.1](#).

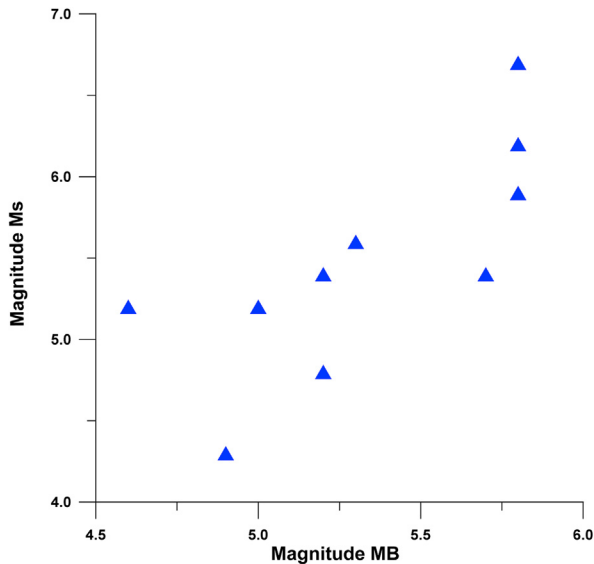


**FIGURE 10.3** Relationship between the magnitudes  $M_s$  and  $m_b$  for volcanic earthquakes. The values of  $M_s$  and  $m_b$  are taken from *Bulletin of ISC*. It is seen in the saturation of magnitude  $m_b$  at value  $m_b = 5.8$ . From Zobin, V.M., 2001. *Seismic hazard of volcanic activity*. *Journal of Volcanology and Geothermal Research* 112, 1–14.

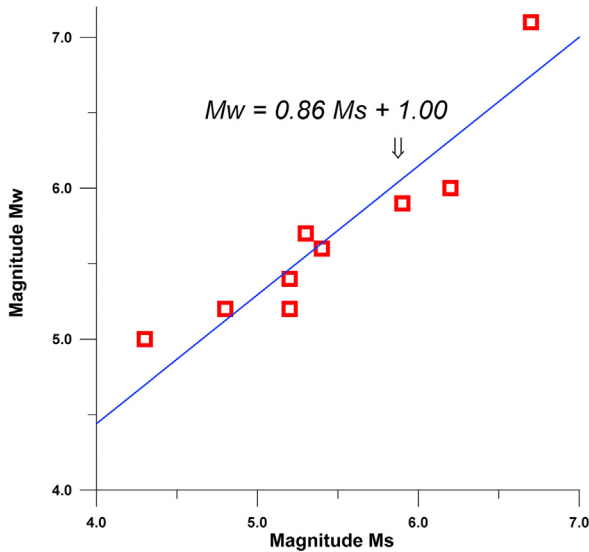


**FIGURE 10.4** Relationship between the magnitudes  $M_s$  and  $M_w$  for volcanic earthquakes. The line of regression is shown. The coefficient of correlation is 0.927. From Zobin, V.M., 2001. Seismic hazard of volcanic activity. *Journal of Volcanology and Geothermal Research* 112, 1–14.

## 10.2 FOCAL RUPTURING OF SIGNIFICANT VOLCANO-TECTONIC EARTHQUAKES AND ITS ROLE IN VOLCANIC PROCESSES

The rupture process of three significant volcano-tectonic earthquakes (Nos. 23, 27, and 28 of Table 10.1) was investigated by Takeo (1992), Zobin and Levina (1998), and Zobin (1999). These reconstructions of the rupture process were realized by the inversion of the near-field strong motion records for Event No. 23 (Teishi Knoll volcano) and the broadband teleseismic  $P$ -waveforms for Event No. 27 (Akademia Nauk volcano) and No. 28 (Grimsvøtn volcano). They were interpreted in terms of barriers or asperities destruction (Ruff, 1983) along the fault plane. The barriers and asperities may be considered as the ancient magmatic bodies that could be destructed by forming the magma intrusion passageway.

### 10.2.1 Rupturing of the Magnitude $M_w$ 5.2 Earthquake Preceding the 1989 Teishi Knoll Submarine Eruption

The 1989 Teishi Knoll earthquake occurred at the final stage of an earthquake swarm 4 days before the eruption (see Section 5.5.1). Its focal mechanism shows that one of its nodal planes coincides with the linear distribution of epicenters of earthquakes following this event (Fig. 10.5). This nodal

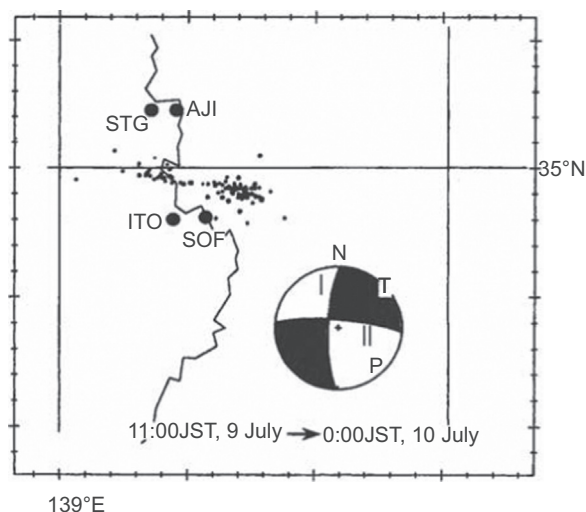


**FIGURE 10.5** Focal mechanism solution for Event 23 and the epicenters of earthquakes that followed this event during 13 h. Large circles show four seismic stations whose records were used in the inversion. The nodal plane II was chosen as the fault plane. From Takeo, M., 1992. The rupture process of the 1989 offshore Ito earthquakes preceding a submarine volcanic eruption. *Journal of Geophysical Research* 97, 6613–6627.

plane striking in the direction of  $N86^{\circ}W$  was taken as the fault plane. Takeo (1992) studied the rupture process of the source of the  $M_w$  5.2 earthquake by using the inversion of digital three-component accelerograms recorded at near seismic stations (Fig. 10.5). Fig. 10.6 shows the comparison of synthetics with records.

Takeo (1992) showed that the  $M_w$  5.2 shock (No. 23) was followed 45 seconds later by the  $M$  4.7 earthquake; he studied these two events together as Event 1 and Event 2 (Fig. 10.7). The rupture of Event 1 initiated in the eastern side subfault as a right-lateral strike-slip motion. Then the rupture stopped due to a barrier; Takeo assumes that this barrier could be an old dyke formation. The excitation of long-period seismic waves stopped for 1.5 seconds after the first stage, but the high-frequency seismic waves excited during the second stage are larger than during other stages.

Therefore it can be suggested that only small-scale fractures took place in the barrier during the second stage. Because the rupture in the third stage involved a little reverse component with the dominant right-lateral strike-slip motion, the barrier seems to have been subjected to vertical shear stress due to the magma intrusion beneath the source zone. The rupture extended westward and the source area of Event 1 reached 4 km in length and 2 km in width. The third stage rupture was prevented by the barrier, which consisted of the old dyke formation existing around the western end of Event 1. This barrier ended the rupture process of Event 1. A new fracture nucleated

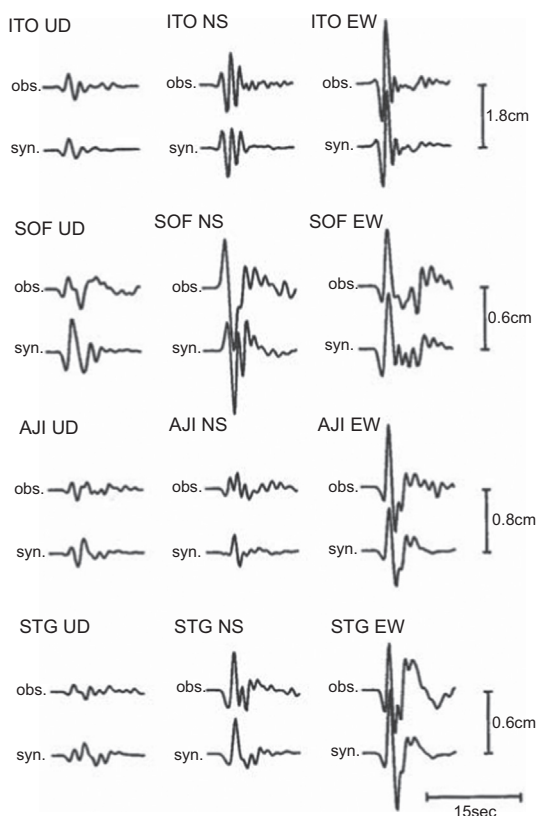


**FIGURE 10.6** Comparison of observed ground displacements of the  $M_w$  5.2 Teishi Knoll earthquake inferred from accelerograms by integrating twice with synthetics *From Takeo, M., 1992. The rupture process of the 1989 offshore Ito earthquakes preceding a submarine volcanic eruption. Journal of Geophysical Research 97, 6613–6627.*

45 seconds later (Event 2) and spread out westward. The rupturing, produced by the  $M_w$  5.2 earthquake, destroyed the old dyke formation and opened the passageway for the magma intrusion.

A high stress concentration on the barrier was recorded due to the magma intrusion. [Takeo \(1992\)](#) assumes that the large stress drop just beneath the Teishi Knoll would have assisted the magma ascent and may have triggered the eruption.

### 10.2.2 Rupturing of the Magnitude $M_w$ 7.1 Earthquake Preceding the 1996 Akademia Nauk Volcano Subaqual Eruption

The phreatomagmatic subaqual eruption in caldera lake of Akademia Nauk volcano, Kamchatka occurred on January 2, 1996, after 4800 years of dormancy. It was preceded by an intensive sequence of earthquakes at depths



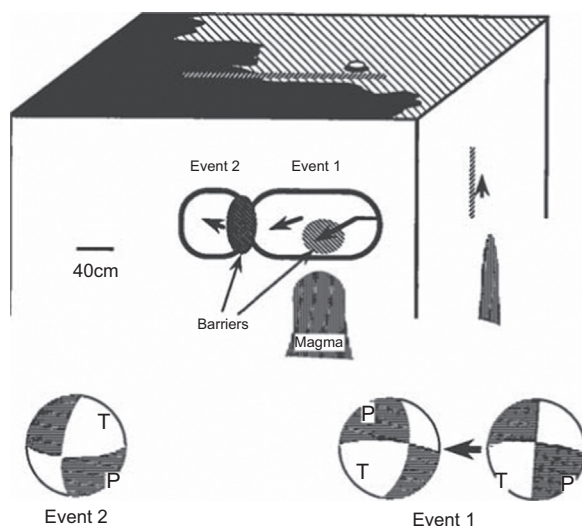


**FIGURE 10.7** Schematic rupture process of the  $M_w$  5.2 earthquake (Event 1) and the following  $M$  4.7 earthquake (Event 2) that preceded the 1989 submarine eruption of Teishi Knoll, Izu Islands *From Takeo, M., 1992. The rupture process of the 1989 offshore Ito earthquakes preceding a submarine volcanic eruption. Journal of Geophysical Research* 97, 6613–6627.

from 0 to 50 km, beginning from January 1, 1996, and including the mainshock with magnitude  $M_w$  7.1 (No. 27 of Table 10.1). The magnitude  $M_w = 7.1$  mainshock was recorded on January 1, about 4 hours after the beginning of the seismic sequence and about 14 hours before the beginning of the eruption. The epicenters of earthquakes preceding the mainshock and the epicenter of the mainshock are shown in Fig. 10.8.

The  $M_w$  7.1 earthquake had a strike-slip nature (Fig. 10.8), with the  $N14^\circ E$  fault plane, oriented along the regional fault connecting the epicenter with the new eruptive site in Karymskoe Lake (Zobin and Levina, 1998). This direction of the fault plane fitted well with the distribution of the epicenters of events that occurred before the mainshock.

A finite-fault, broadband teleseismic  $P$ -waveform inversion (Fig. 10.9) allowed the reconstruction of a dislocation model of the faulting process. The rupture model is presented in terms of a contour map of the final slip of the fault (Fig. 10.10). Two main asperities, outlined by the slip contour of 125 cm, were destroyed; one at depths between 0 and 15 km and another at depths between 25 and 35 km. The majority of small earthquake foci were located away from the maximum asperities and filled a space between two zones of asperities at depths from 12 to 20 km. Probably, this large seismic event represented two shocks, one shallow and another at a depth of about 30 km (Zobin and Levina, 1998).

The great amount of juvenile calc-alkaline basalt, representing about 95% of the ejected material (Belousov and Belousova, 2001) supports the

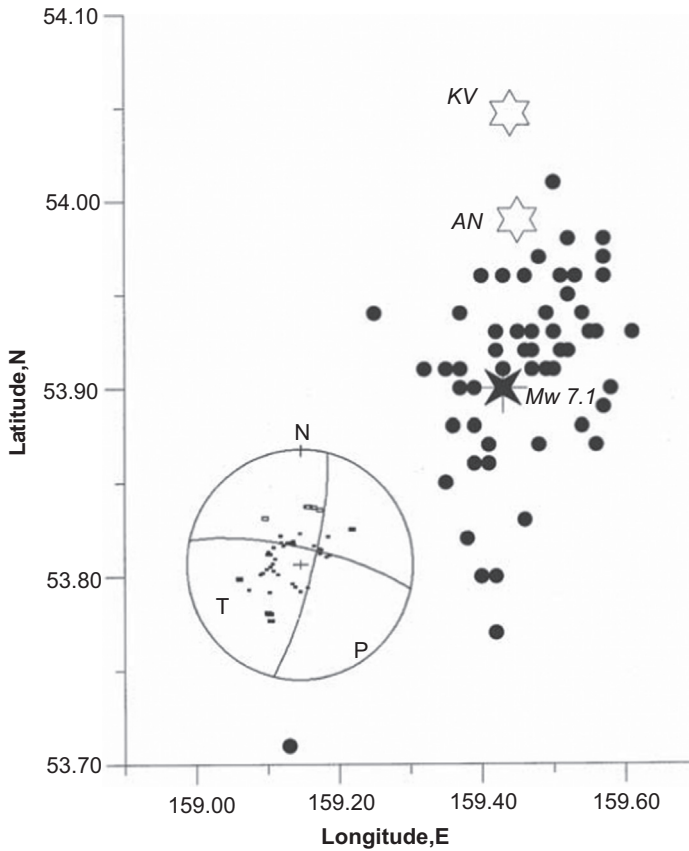


**FIGURE 10.8** Seismicity preceding the appearance of the  $M_w = 7.1$  earthquake related to the eruption of Akademia Nauk volcano (No. 27). The foreshocks that were recorded from 06:20 to 09:57 of January 1, 1996, are shown as filled circles. The volcanoes Karymsky (KV) and Akademia Nauk (AN) are shown as open stars. The epicenter of the  $M_w = 7.1$  earthquake is shown by the filled star. The focal mechanism solution for this event is shown in the left corner. The focal mechanism was published by [Zobin and Levina \(1998\)](#). Data from the Kamchatka Regional Earthquake Catalog.

inference of the deeper seismic source. It seems possible that during the large earthquake an ancient deep-seated magmatic column was destroyed and a new system of faults was formed that would act as a magma passageway.

### 10.2.3 Rupturing of the Magnitude $M_w$ 5.6 Earthquake Preceding the 1996 Grimsvøtn Volcano Subglacial Eruption

The 1996 Grimsvøtn earthquake (No. 28 from [Table 10.1](#)) occurred as the mainshock of the preceding earthquake sequence. It occurred 36 hours before the eruption (see Section 2.3.1). The earthquake had an inverse-type nature,

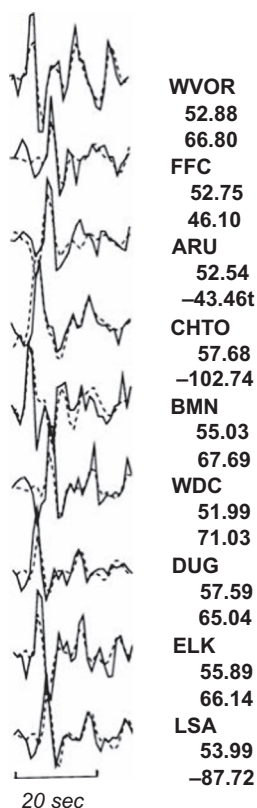


**FIGURE 10.9** The comparison of observed (solid line) and synthetic (dashed line)  $P$ -waveforms of the  $M_w$  7.1 earthquake of January 1, 1996, preceding the subaqual Akademia Nauk volcano eruption. The epicentral distance (in degrees) and azimuth from the epicenter (in degrees) are indicated for each seismogram pair. From Zobin, V.M., Levina, V.I., 1998. Rupture history of the 1996 January 1  $M_s$  6.6 volcanic earthquake preceding the simultaneous eruption of Karymsky and Akademia Nauk volcanoes in Kamchatka, Russia. *Journal of Geophysical Research* 103, 18315–18324.

with the submeridional fault planes (Harvard..., 1977–1999). This direction of the fault planes fitted well with the distribution of the epicenters of events that occurred just after the mainshock (Fig. 10.11).

A finite-fault, broadband teleseismic  $P$ -waveform inversion allowed the reconstruction of a dislocation model of the faulting process. The rupture model is presented in terms of a contour map of the final slip on the fault (Fig. 10.12).

The inversion showed that the rupture had developed downdip from the mainshock hypocenter and the main asperity was ruptured at depths between 4 and 6 km. A comparison with the deep distribution of aftershocks showed that their hypocenters were situated above the destroyed asperities at depths between 0 and 4 km. These space-time seismic patterns and the results of the

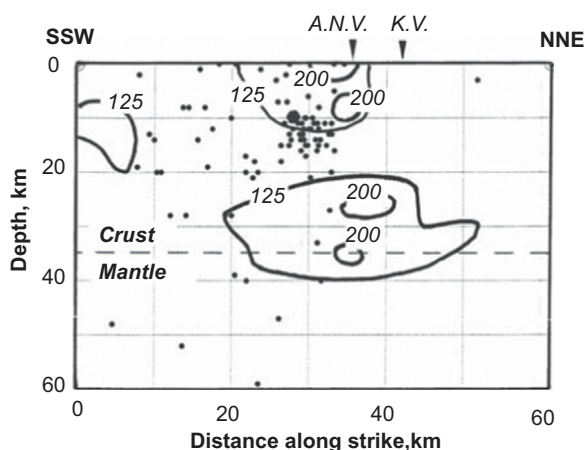


**FIGURE 10.10** The comparison of the dislocation model, obtained for the  $M_w$  7.1 earthquake preceding the January 1996 Akademia Nauk volcano eruption, with seismic and volcanic characteristics. Solid circles denote the hypocenters of earthquakes with magnitude  $M > 3.0$  recorded before the beginning of the eruption. The hypocenter of the mainshock is indicated by a solid hexagon. Cumulative slip is contoured at 75-cm intervals beginning from 125 cm. ANV, Akademia Nauk volcano; KV, Karymsky volcano. The border between the crust and mantle is shown by the dashed line. From Zobin, V.M., Levina, V.I., 1998. Rupture history of the 1996 January 1  $M_s$  6.6 volcanic earthquake preceding the simultaneous eruption of Karymsky and Akademia Nauk volcanoes in Kamchatka, Russia. *Journal of Geophysical Research* 103, 18315–18324.

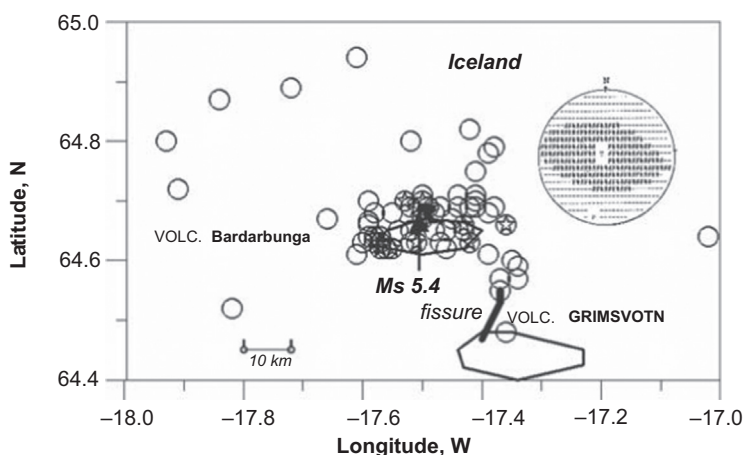


**FIGURE 10.11** The September 29, 1996 mainshock epicenter (solid triangle), its focal mechanism solution and the epicenters of its aftershocks (open circles), which preceded and accompanied the subglacial eruption of Grimsvötn volcano, Iceland. The linear distribution of the aftershocks that occurred during first 19 h of activity coincides with the directions of two nodal planes of the focal mechanism solution. Focal mechanism solution is taken from Harvard... (1977–99). The black heavy line shows the volcanic fissure, which was formed during the eruption. The volcano calderas are shown by heavy-line ovals. From Zobin, V.M., 1999. The fault nature of the  $M_s$  5.4 volcanic earthquake preceding the 1996 subglacial eruption of Grimsvötn volcano, Iceland. *Journal of Volcanology and Geothermal Research* 92, 349–358.

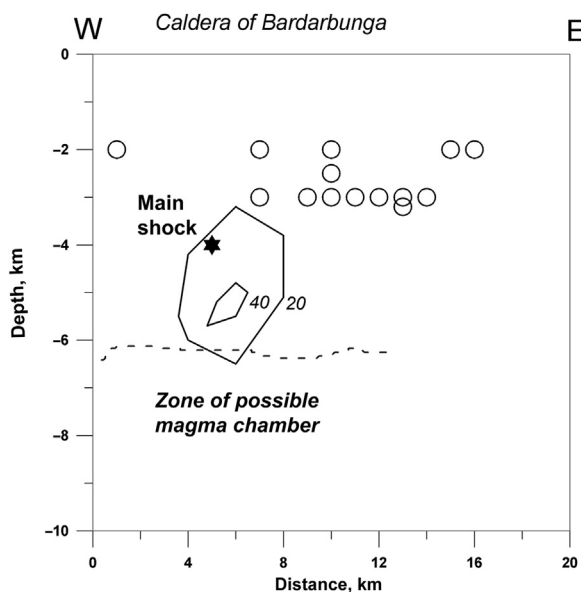


**FIGURE 10.12** Cross-section beneath the caldera of Bardarbunga. The circles are hypocenters of the aftershocks ( $M_L \geq 3.0$ ); the star is the hypocenter of the mainshock. The slip contours are shown for 20 and 40 cm and outline the asperity broken during the rupturing of the mainshock fault. The zone of a possible magma chamber is shown below the asperities. *From Zobin, V.M., 1999. The fault nature of the  $M_s$  5.4 volcanic earthquake preceding the 1996 subglacial eruption of Grimsvötn volcano, Iceland. *Journal of Volcanology and Geothermal Research* 92, 349–358.*

finite-fault inversion may be interpreted as follows: There was a magma chamber under Bárðarbunga volcano covered by strong rocks or asperities. The mainshock of September 29 broke the asperity and formed a path for magma transport. Then a fissure formed connecting the opened magma chamber with the feeding structures of Grimsvötn volcano, and magma was injected laterally into the site of the eruption. The continuous seismic activity during the third period was related to the collapse of rocks around the released parts of the magma chamber. The role of the  $M_w$  5.6 earthquake in this sequence was decisive.

### 10.3 THE MAGNITUDE 7 VOLCANO-TECTONIC EARTHQUAKES IN VOLCANIC PROCESSES

For more ancient large volcano-tectonic earthquakes, we had no rupture reconstructions. At the same time, their participation in eruptive processes is of interest. Here the author discusses two of them, the largest events with magnitude  $M_w = 7.0$  (Nos. 2 and 3 of [Table 10.1](#)).

### 10.3.1 Event No. 2, Katmai, Alaska

The 1912 Katmai eruption was the most voluminous eruption of the 20th century, producing about  $30 \text{ km}^3$  of products. The formerly cone-shaped summit of Mount Katmai vanished, leaving in its place a caldera 3 km in diameter and nearly 1 km deep. The collapse was apparently caused by the withdrawal of supporting magma from beneath Mount Katmai toward Novarupta, the actual vent of the eruption, 10 km west of Mount Katmai (Abe, 1992). There was no information about the previous activity of Katmai in historical times.

The seismic activity was recorded beginning about 5 days before the eruption, but the strongest earthquakes were observed during the eruptive activity. The eruption began on June 5, 1912, with frequent minor explosions. Then, a series of big explosions was observed from 13 hours of June 6 to July 8. The eruption continued until July 21 (Simkin and Siebert, 1994), but without significant explosions. The sequence of strong earthquakes ( $M_s \geq 6.0$ ) was recorded beginning from the second tremendous explosion of June 6, and the largest event with magnitude  $M_s = 7.0$  occurred at 21:26 (local time) on June 7, at the final part of the violent phase of the eruption, when from 21 to 22 hours, 42% of the total seismic energy was released (Abe, 1992).

Abe (1992) considers that a series of substantial collapses of Mount Katmai occurred during this seismic climax. Therefore it is speculated that the  $M_w = 7.0$  seismic event, which was recorded at the seismic climax and before the third last great explosion, was the decisive one in the Mount Katmai caldera collapse, opening the fault as a passageway for the withdrawal of supporting magma from beneath Mount Katmai toward Novarupta.

### 10.3.2 Event No. 3, Sakurajima, Japan

The 1914 Sakurajima eruption was also one of the largest eruptions of the century (see Section 6.4.1). The eruption began on January 12 at about 08 hours (local time) with the emissions of white smoke at the South Peak, and 2 hours after explosions occurred at the western and eastern slopes, where vents had opened. Just after midnight of January 13, lava began to effuse from the craterlets on both the eastern and western slopes (Yokoyama, 1997). It was the first flank eruption from vents situated on the western and eastern flanks of the volcano in historical times. There were recorded earlier only the eruptions from the summit crater Minami-dake or from the flank vents situated on the NE, SW, or south slopes of the volcano (Simkin and Siebert, 1994).

Seismic activity near the volcano was recorded beginning from January 9. Earthquakes increased in number and magnitude, with the maximum activity on January 11. After the opening of vents in the morning of January 12, earthquake activity rapidly decreased. Then, at 18:30 on January 12, the

$M_w = 7.0$  shock occurred. About 26 hours later, the great lava extrusion began (Yokoyama, 1997). Therefore it was speculated that this large earthquake opened the passageway for magma and was decisive in the development of eruption from rather calm flank explosions to a greater volume lava eruption.

The analysis of the five largest volcanic earthquakes shows that all events were associated with the eruptions of quasi-dormant volcanoes (Katmai and Akademia Nauk) or with a new type of activity for historical time (Teishi Knoll, Grimsvøtn, and Sakurajima). The fault nature (events Nos. 23, 27, and 28) and the temporal position of the events in the eruption process allow us to speculate that these earthquakes may have formed new passageways for magma movement.

## 10.4 SEISMIC HAZARD OF SIGNIFICANT VOLCANO-TECTONIC EARTHQUAKES

The traditional estimation of seismic hazard includes the following: (1) the estimation of maximum magnitude  $M_{\max}$ , (2) its recurrence time, and (3) the law of earthquake intensity decrease with distance for a variety of magnitudes (e.g., Basham et al., 1982). We had no possibility to have a long-term sequence of volcano-tectonic earthquakes for a volcano. So we discuss the seismic hazard for a type of eruption. Table 10.2 describes the types of eruptions and the position of the seismic event in the eruptive process. Classification of 27 of a total of 32 eruptions was taken from the catalog of volcanoes of the world by Simkin and Siebert (1994) and GVN (2011). The position of the largest seismic event in the eruptive process was estimated from publications about these eruptions.

Significant earthquakes that occurred near volcanoes but without relation to eruptive activity (Kubotera, 1988; Abe, 1979; McNutt, 1996) were not included in this study. Among them, were the  $M_s$  6.3 of 1916 earthquake that occurred near Asama volcano and the  $M_s$  6.1 of 1975 earthquake that occurred to the north of Aso caldera. There was a problem with the magnitudes estimated by near stations for rather small events, since, as a rule, they were overestimated. For example, the magnitude of the earthquake of July 7, 1990, related to the eruption of Unzen volcano, was determined by near stations as  $M$  4.8 (Nakada et al., 1999). At the same time, ISC gave the magnitude  $mb$  4.4 and no estimation of  $M_s$ . These doubtful events were also not included in our catalog. The 1975  $M_s$  7.1 Kalapana, Hawaii, earthquake triggered the very small summit eruption of Kilauea (Klein et al., 1987). This eruption was very insignificant and short; the real relation between the earthquake and the eruption is unclear. McNutt (1996) did not include it in his list of large seismic events near volcanoes but only mentioned it among the large regional earthquakes that could trigger the eruptions. This earthquake was not included in Table 10.1.

**TABLE 10.2** Short Volcanological Information About the Significant Earthquakes (SE) of Table 10.1.

| No. | Volcano,<br>Year      | Type of<br>Eruption | $M_w$<br>of SE | Before<br>Eruption | Initial<br>Stage | Paroxysmal<br>Stage | Final<br>Stage | $T$ (years), Remarks                                    |
|-----|-----------------------|---------------------|----------------|--------------------|------------------|---------------------|----------------|---------------------------------------------------------|
| 1   | Usu, 1910             | 3                   | 5.4            | *1                 |                  |                     |                | North flank, $T = \infty$                               |
| 2   | Katmai, 1912          | 1a                  | 7.0            |                    | *2               |                     |                | $T = \infty$                                            |
| 3   | Sakurajima,<br>1914   | 2,b, 3              | 7.0            |                    | *3               |                     |                | The west and east flanks, $T = \infty$                  |
| 4   | Paricutin,<br>1943    | 1, 2a, 4            | 4.9            | *4                 |                  |                     |                | New volcano, $T = \infty$                               |
| 5   | Bezymianny,<br>1956   | 1                   | 5.3            |                    |                  | *5                  |                | Directed blast, $T = \infty$                            |
| 6   | Miyakejima,<br>1962   | 2a, 3               | 6.1            |                    |                  |                     | *6             | The NE flank eruption with radial fissure, $T = \infty$ |
| 7   | Sheveluch,<br>1964    | 1                   | 5.2            |                    |                  | *7                  |                | Directed blast, $T = 110^{29}$                          |
| 8   | Fernandina,<br>1968   | 1a                  | 5.4            |                    |                  | *8                  |                | Caldera collapse, $T = \infty$                          |
| 9   | Arenal, 1968          | 1, 2 <sup>a</sup>   | 5.1            | *9                 |                  |                     |                | The west flank and summit, $T = \infty$                 |
| 10  | Beerenberg,<br>1970   | 1, 2a, 4            | 5.7            | *10                |                  |                     |                | The NE and SW flanks and summit, $T = \infty$           |
| 11  | Curacoa<br>Reef, 1973 | 5                   | 5.9            | *11                |                  |                     |                | Submarine, $T = \infty$                                 |
| 12  | Tiatia, 1973          | 2b                  | 4.6            | *12                |                  |                     |                | $T = \infty$                                            |
| 13  | Etna, 1974            | 3                   | 4.7            | *13                |                  |                     |                | West flank, $T = 211$                                   |

*(Continued)*



**TABLE 10.2 (Continued)**

| No. | Volcano,<br>Year      | Type of<br>Eruption | $M_w$<br>of SE | Before<br>Eruption | Initial<br>Stage | Paroxysmal<br>Stage | Final<br>Stage | $T$ (years), Remarks                                                                             |
|-----|-----------------------|---------------------|----------------|--------------------|------------------|---------------------|----------------|--------------------------------------------------------------------------------------------------|
| 14  | Tolbachik,<br>1975    | 1, 2a, 4            | 4.9            | *14                |                  |                     |                | New volcano, $T = \infty$                                                                        |
| 15  | Nyiragongo,<br>1977   | 1, 3                | 5.1            | *15                |                  |                     |                | The simultaneous eruption of N, S, and W flanks,<br>$T = \infty$                                 |
| 16  | Usu, 1977             | 1                   | 4.7            |                    |                  |                     | *16            | $T = 124$                                                                                        |
| 17  | St. Helens,<br>1980   | 1, 2a               | 5.4            |                    |                  | *17                 |                | Directed blast, $T = \infty$                                                                     |
| 18  | Una Una,<br>1983      | 1                   | 4.9            |                    | *18              |                     |                | $T = 45$                                                                                         |
| 19  | Miyakejima,<br>1983   | 2a, 3               | 6.0            |                    |                  |                     | *19            | SW flank, $T = 271$                                                                              |
| 20  | Beerenberg,<br>1985   | 2a, 4               | 5.0            |                    | *20              |                     |                | NE flank, $T = 12$                                                                               |
| 21  | Oshima,<br>1986       | 1, 3                | 5.9            |                    |                  |                     | *21            | The eruption of central crater with formation of<br>fissure in N and NW of caldera, $T = \infty$ |
| 22  | Lonquimay,<br>1988    | 2a, 3               | 5.2            |                    | *22              |                     |                | The flank crater with radial fissure, $T = \infty$                                               |
| 23  | Teishi Knoll,<br>1989 | 5                   | 5.2            | *23                |                  |                     |                | Submarine eruption, $T = \infty$                                                                 |
| 24  | Pinatubo,<br>1991     | 1, 2a               | 5.6            |                    |                  | *24                 |                | $T = \infty$                                                                                     |

|    |                     |                     |     |     |     |                                                                                  |
|----|---------------------|---------------------|-----|-----|-----|----------------------------------------------------------------------------------|
| 25 | Hudson, 1991        | 1, 3                | 5.8 |     | *24 | The radial fissure, $T = \infty$                                                 |
| 26 | Etna, 1991          | 3                   | 4.7 |     | *26 | SE flank, $T = 23$                                                               |
| 27 | Akademia Nauk, 1996 | 4 <sup>30</sup>     | 7.1 | *27 |     | Subaqual caldera eruption, $T = \infty$                                          |
| 28 | Grimsvötn, 1996     | 2a, 4 <sup>28</sup> | 5.6 | *28 |     | Subglacial eruption, the fissure eruption on the N flank, $T = 58$ <sup>31</sup> |
| 29 | Usu, 2000           | 2a <sup>32</sup>    | 4.9 |     | *29 | NW flank, $T = \infty$                                                           |
| 30 | Miyakejima, 2000    | 2a, 5 <sup>33</sup> | 6.2 |     | *30 | Submarine W flank eruption, $T = 165$                                            |
| 31 | Miyakejima, 2000    | 1a <sup>34</sup>    | 6.4 |     | *31 | Caldera collapse, $T = \infty$ <sup>34</sup>                                     |
| 32 | Dabbahu, 2005       | 4 <sup>35</sup>     | 5.5 | *35 |     | Rift intrusion, $T = \infty$ <sup>35</sup>                                       |

*Notes:* Type of eruption: 1, central eruption; 1a, caldera collapse; 2a, one-side flank eruption; 2b, two-side flank eruption; 3, radial fissure; 4, regional fissure; 5, submarine eruption. T (in years) is the recurrence time. The superscripts denote the following references: 1, Okada, 1982; 2, [Abe, 1992](#); 3, [Yokoyama, 1997](#); 4, Yokoyama and de la Cruz-Reyna, 1990; 5, Gorshkov and Bogoyavlenskaya, 1965; 6, Minakami, 1964; 7, Tokarev, 1981; 8, Filson et al., 1973; 9, Minakami et al., 1969; 10, Siggerud, 1972; 11, Gibowicz et al., 1974; 12, Zobin, 1976; 13, Bottari et al., 1975; 14, Zobin, 1979a,b; 15, Hamaguchi et al., 1992; 16, Wano and Okada, 1980; 17, Lipman and Mullineaux, 1982; 18, USGS; 19, Tada and Nakamura, 1988; 20, Havskov and Atakan, 1991; 21, Yamaoka et al., 1988; 22, Barrientos and Acevedo-Aranguiz, 1992; 23, Matsumura et al., 1991; 24, Harlow et al., 1996; 25, Gonzalez-Ferran, 1995; 26, Ferrucci and Patane, 1993; 27, [Zobin and Levina, 1998](#); 28, BGVN, 21–09, 1996; 29, Melekestsev et al., 1991; 30, Belousova, 1999, pers. com.; 31, Gudmundsson, 1996; 32, Zobin et al., 2005a,b; 33, Uhira et al., 2005; 34, Geshi et al., 2002; 35, Ayele et al., 2007.

*Source:* The data about eruptions are taken from Simkin, T., Siebert, L., 1994. Volcanoes of the World, second ed. Geoscience Press, Tucson. p. 349, except for any specially marked cases.

The data for recurrence time were taken from the catalog by [Simkin and Siebert \(1994\)](#). The data for calculation of the law of decreasing intensity with distance were taken for three Japanese earthquakes from *Bulletin ISC* ([Table 10.3](#)).

For all regressions calculated in this study, the correlation was significant at the 95% confidence level.

#### 10.4.1 Maximum Magnitude $M_{\max}$

The problem of maximum earthquake magnitude for any region or local zone is very important in the evaluation of seismic hazards. The size of the maximum magnitude is decisive in the prediction of the destructive effects of seismic activity. The maximum magnitude may be estimated: (1) by consideration of the maximum fault area that can break in a single event; (2) by the magnitude truncation in the observed seismicity for source zones in which the return period for the maximum magnitude is shorter than the observation period (e.g., [Basham et al., 1982](#)); or (3) by statistical study of the catalog (e.g., [Kagan, 1997](#)). However, in our case, these types of evidence are not available. We work not with a single volcano but with a group of volcanoes all over the world. Therefore we estimate the maximum magnitude  $M_{\max}$  for this set of volcanoes combining them in two ways, based on the data of [Table 10.1](#) and [Table 10.2](#), for (1) the type of eruption and (2) the stage of eruption.

The seismicity of volcanoes depends strongly on the type of eruption. Central and flank eruptions are characterized by different development of the swarms of earthquakes ([Gorelchik et al., 1990](#); [Zobin, 1988](#)). Earthquake swarms, observed at Kamchatkan volcanoes and preceding central eruptions, are characterized by continuously increasing numbers and magnitudes of earthquakes up to the beginning of the eruption. Earthquake swarms preceding flank eruptions are characterized by the presence of a maximum number and magnitude of earthquakes at the beginning of the swarm, and a decrease of both just before the eruption.

[Gresta et al. \(1990\)](#) showed that the eastern and western flanks of Mount Etna are characterized by different morphological, structural, volcanic, and seismic data. They recognized the higher “degree” of tectonization on the eastern slope from the morphological and structural data. The eastern slope also has shallower and more intensive seismicity than the western slope. They attribute these differences to higher extensional stresses on the eastern slope.

[Fig. 10.13](#) demonstrates the tendency of the change of the maximum magnitude in volcanic earthquake swarms, preceding the flank eruptions with the altitude of the newborn cones, of the 4750-m height Klyuchevskoy volcano, Kamchatka, in 1953–87. The values of  $M_{\max}$  of swarms preceding the flank eruptions at lower altitudes (1360–1900 m) were generally greater than the  $M_{\max}$  of swarms preceding the birth of new cones at higher altitudes (2300–3340 m). This tendency may depend on at least two factors: (1) the distance between the new flank cone and the central conduit of the volcano and (2) the mechanical quality of the rocks where the earthquake rupture was formed. Both factors could

**TABLE 10.3** Macroseismic Data for Three Volcano-Tectonic Earthquakes.

| Site       | $I_{JMA}$ | Event no. 19 $I_{MM}$ | $R$ (km) | $I_{JMA}$ | Event no. 21 $I_{MM}$ | $R$ (km) | $I_{JMA}$ | Event no. 23 $I_{MM}$ | $R$ (km) |
|------------|-----------|-----------------------|----------|-----------|-----------------------|----------|-----------|-----------------------|----------|
| Miyakejima | 5         | 7                     | 17       | 4         | 6                     | 66       |           |                       |          |
| Oshima     | 3         | 5                     | 85       |           |                       |          | 3         | 5                     | 36       |
| Hacijojima | 3         | 5                     | 102      |           |                       |          |           |                       |          |
| Tateyama   | 3         | 5                     | 109      | 4         | 6                     | 46       |           |                       |          |
| Niijima    | 3         | 5                     | 45       | 4         | 6                     | 43       |           |                       |          |
| Ajiro      | 2         | 4                     | 112      | 3         | 5                     | 44       | 4         | 6                     | 6        |
| Katsuura   | 2         | 4                     | 143      | 2         | 4                     | 102      | 2         | 4                     | 114      |
| Yokohama   | 2         | 4                     | 152      | 3         | 5                     | 85       | 3         | 5                     | 72       |
| Tokyo      | 2         | 4                     | 180      | 3         | 5                     | 113      | 3         | 5                     | 42       |
| Irozaki    |           |                       |          | 3         | 5                     | 51       | 3         | 5                     | 42       |
| Kofu       |           |                       |          | 3         | 5                     | 127      | 2         | 4                     | 92       |
| Mishima    |           |                       |          | 2         | 4                     | 61       | 3         | 5                     | 23       |
| Shizuoka   |           |                       |          | 2         | 4                     | 94       | 2         | 4                     | 71       |
| Chiba      |           |                       |          | 2         | 4                     | 125      | 2         | 4                     | 117      |

Notes: Number of events correspond to [Table 10.1](#).  $I_{JMA}$  is the earthquake intensity of Japan Meteorological Agency scale;  $I_{MM}$  is the earthquake intensity of the Modified Mercalli scale; and  $R$  is the hypocentral distance.

Source: Data were taken from the Bulletin of ISC.

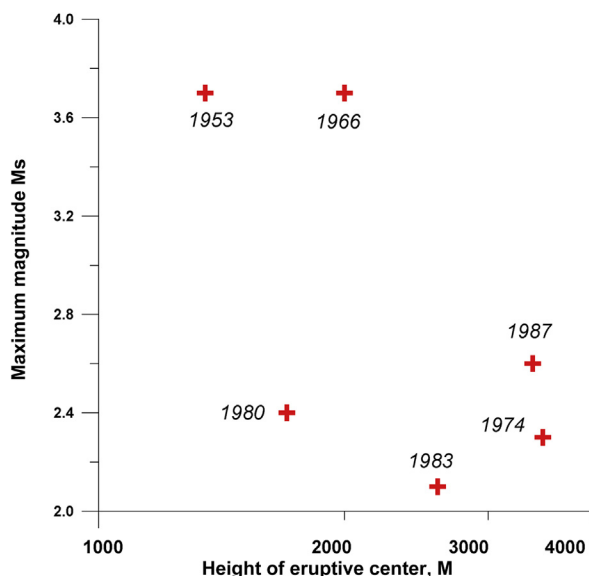


**FIGURE 10.13** Dependence of maximum magnitude of volcanic earthquakes on the eruption site altitude for Klyuchevskoy volcano. The mean altitudes of flank eruption sites were taken from Khrenov et al. (1991). The magnitudes  $M_s$  were calculated from the energy class  $K_s$  (Zobin, 1996). The values of energy class  $K_s$  were taken from Tokarev (1981), Zobin (1979a), and Gorelchik (1989; personal communication, 1999). The numbers near crosses indicate the date of the flank eruption. From Zobin, V.M., 2001. *Seismic hazard of volcanic activity. Journal of Volcanology and Geothermal Research* 112, 1–14.

constrain the length of the earthquake rupture and the size of  $M_{\max}$ . The first factor, that is the decrease of the distance between the central conduit and the new eruptive site with the increase of the altitude of the site, would be decisive in the length of rupture for producing the earthquake with  $M_{\max}$ . At the same time, the decreasing of mechanical properties of the rocks at high altitudes could prevent large ruptures compared to those at lower altitudes. This example shows that the conditions of forming an eruptive site could be an important factor in the size of the maximum seismic event related to the eruption.

Therefore we divided our data into five groups: earthquakes related to (1) central eruption, (2) caldera collapse, (3) central eruption with flank or fissure eruption, (4) flank and/or fissure eruption without central eruption, and (5) submarine eruptions.

Fig. 10.14 shows the magnitude distribution for the largest volcanic earthquakes associated with these five groups of eruptions. The  $M_{\max}$  was taken as the maximum magnitude of earthquake recorded in the century for each type of eruption. Due to this distribution, central eruptions are characterized by the appearance of earthquakes of  $M_{\max}$  of 5.4. It is the smallest  $M_{\max}$  value among all types of eruptions. The central eruptions followed by the

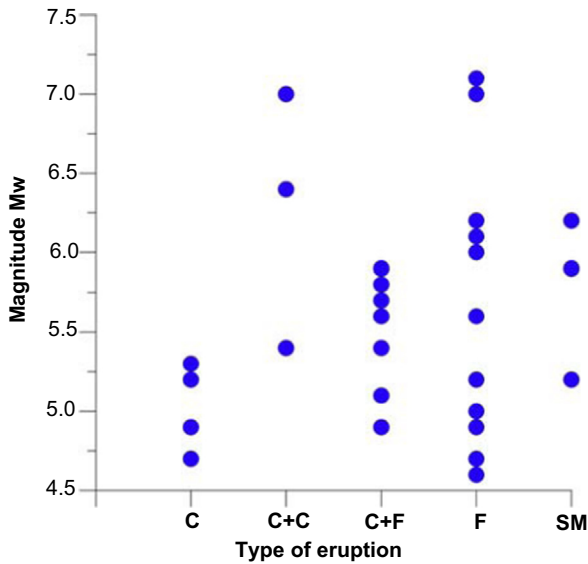


**FIGURE 10.14** Distribution of magnitudes  $M_w$  for five types of volcanic eruptions. C, Central eruption; C + C, caldera collapse; C + F, the central eruption with flank and/or fissure eruption; F, the flank and/or fissure eruption; SM, the submarine eruption. Data from Table 10.2.

appearance of flank and/or fissure eruptive centers are characterized by slightly higher  $M_{\max}$  of the earthquake (5.9).

The earthquakes with the largest  $M_{\max}$  were recorded for the caldera collapses (7.0) and the flank and/or fissure eruptions without central activity (7.1). The  $M_{\max}$  for submarine eruptions was equal to 6.2.

The next set of  $M_{\max}$  values was obtained for the earthquakes occurring at different stages of eruptions (Fig. 10.15). We can see that the largest volcanic earthquakes occur at the preliminary or initial stages of eruption ( $M_{\max}$  7.1 and 7.0, respectively). The paroxysmal stage is characterized by relatively low-magnitude earthquakes (5.6), and at the final stage of eruption rather strong earthquakes can be observed ( $M_{\max}$  6.4).

Cubellis et al. (2007) estimated the maximum magnitude for the volcano-tectonic earthquakes occurring at Vesuvius volcano using the 30-year seismic observations around the volcano as well as the historical data. Comparison of instrumental seismicity and historical data reveals two significantly different energy levels: a lower earthquake level with  $M_{\max} = 4.5$ , corresponding to current seismicity and that which accompanied volcanic activity in the eruptive period from 1631 to 1944; and an upper earthquake level with  $M_{\max} = 5.4$ , represented by the AD 62 earthquake. The authors suggest that these two levels correspond to two stress states and different seismogenic structures. Considering the central eruptions as characteristic of Vesuvius, these estimations of  $M_{\max}$  coincide with ours ( $M_{\max}$  5.4) suggested for central eruptions.

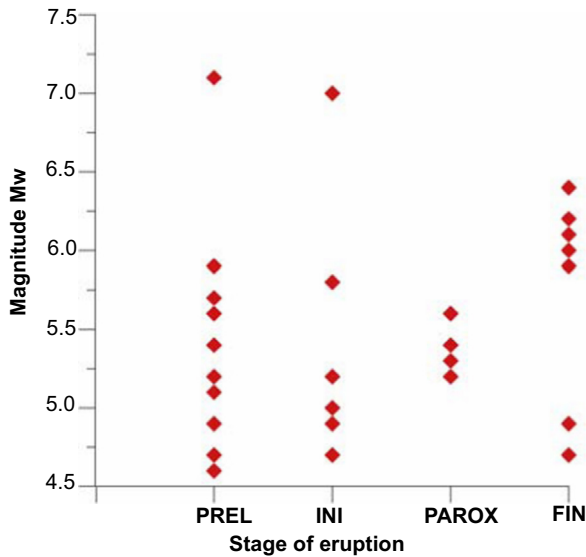


**FIGURE 10.15** Distribution of magnitudes  $M_w$  for four stages of eruptions. *PREL*, the preliminary stage; *INI*, the initial stage; *PAROX*, the paroxysmal stage; *FIN*, the final stage. Data from Table 10.2.

The problem of the maximum magnitude of volcanic earthquakes was discussed also by [Okada \(1983\)](#). He examined the earthquakes related to “direct blast,” or “sector collapse” eruptions of Unzen (1792), Bandai-san (1888), Bezymianny (1956), Sheveluch (1964), and Mount St. Helens (1980). In addition to the data listed in Table 10.1, [Okada \(1983\)](#) estimated the magnitude between 5 and 5.5 for the earthquake related to the Bandai-san eruption, according to its macroseismic effect and to its record in Tokyo (epicentral distance 220 km), with amplitude of 0.2 mm on the Gray-Milne seismograph. He also considered, according to macroseismic data, the magnitude 5–5.5 for the earthquake that accompanied the 1792 explosion of Unzen. Therefore he concluded that there tends to exist a maximum limit in magnitude at around 5, except for those in which regional tectonics are required for their direct causes.

[Benoit \(1998\)](#) has studied the dependence of the  $M_{\max}$  of volcanic earthquakes upon the type of eruption and the Volcanic Explosivity Index (VEI). He analyzed 180 swarms of earthquakes that occurred near volcanoes, mainly from 1979 to 1989. The  $M_{\max}$  values ranged from 0.5 to 6.2. The relations between  $M_{\max}$  and VEI did not demonstrate any significant correlation. At the same time, [Benoit \(1998\)](#) showed that the eccentric and radial fissure eruptions were preceded by larger shocks than central vent eruptions. This result is in good coincidence with ours.

The reason for the significant difference between the  $M_{\max}$  of earthquakes associated with the central eruptions and of earthquakes related to more complex eruptions, such as forming of regional or flank fissures, may be the

difference in magma transport (Zobin, 1988). It was suggested that large earthquakes formed new fissures for magma discharge in zones of regional faults. The “renovation” of the volcano conduit before and during the central eruption does not require such a large effort. For central eruptions, we suggest that the mechanism of failure is a result of local dyke intrusion within the volcanic edifice, which can involve magmatic driving forces and fluid pressurization mechanisms (e.g., Elsworth and Voight, 1995). This mechanism could trigger moderate-size earthquakes, whose magnitudes would be limited by the size of volcanic edifice and which could be accompanied by a failure of volcanic flanks or sector collapses.

#### 10.4.2 Attenuation of Earthquake Intensity with Distance for Volcanic Earthquakes

The *Bulletin of ISC* published the macroseismic data for three events of magnitude interval 5.2–6.0 (Nos. 19, 21, and 23 from Table 10.1) which were related to the eruptions of Izu Islands, Japan, volcanoes. They are collected in Table 10.3. These data were prepared by the Japan Meteorological Survey (JMA) according to the JMA macroseismic scale and then were transformed into a more popular MM macroseismic scale using the comparative table published by Tiedemann (1992). The intensity–distance equation for 31 points shown in Fig. 10.16 was calculated in the following form:

$$I = aM_s - \log R - cR + d \quad (10.2)$$

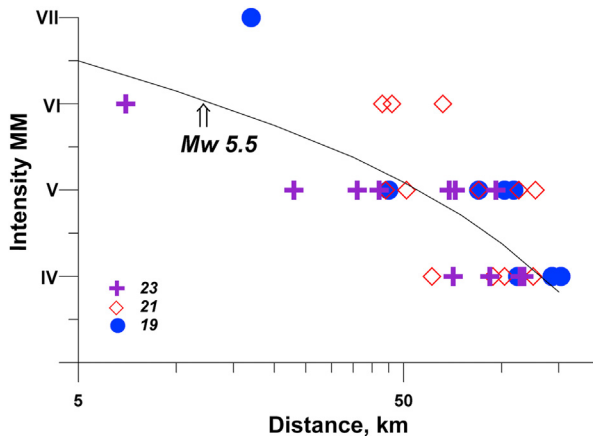


**FIGURE 10.16** Decrease of earthquake intensity of volcano-tectonic earthquakes with distance. The curve calculated from Eq. (10.3) for the earthquake with magnitude  $M_w$  5.5, focal depth 0 km, is shown. The data are marked by different symbols for three earthquakes, Nos. 19, 21, and 23 from Table 10.1. Intensity data were taken from *Bulletin of ISC*. From Zobin, V.M., 2001. *Seismic hazard of volcanic activity. Journal of Volcanology and Geothermal Research* 112, 1–14.



where  $R = (\Delta^2 + h^2)^{1/2}$  is the hypocentral distance in km,  $\Delta$  is the epicentral distance in km,  $h$  is the focal depth in km, the coefficients  $b$  and  $c$  are the coefficients of geometric spreading of seismic waves and absorption, respectively, and  $d$  is a free member. This form of the intensity–distance equation was recommended for macroseismic studies by the New IASPEI Manual of Seismological Observatory Practice (Bormann and Bergman, 2000) and allows the expression of the intensity attenuation as a function of magnitude and distance, as well as the estimation of the inelastic properties of rocks in the region.

The following law for earthquake intensity MM attenuation with epicentral distances  $>5$  km was obtained:

$$I = 0.66M_w - 1.13\log R - 0.0072R + 3.73 \quad (10.3)$$

The curve, calculated according to Eq. (10.3) for the  $M_w$  5.5 earthquake, which occurred at a focal depth of 0 km just below the volcanic crater, is shown in Fig. 10.16 and can predict a possible macroseismic effect.

At the same time, the epicenter of a significant volcanic earthquake may not coincide with the volcanic crater position. Zobin (1996) showed that the position of the largest earthquake depends strongly on the crustal thickness  $H$  beneath the volcano. A sketch of Fig. 10.17 illustrates the situation for continental and oceanic volcanoes. The epicentral area of volcanic earthquakes is the surface projection of the faulted volume around the magma conduit. For any given

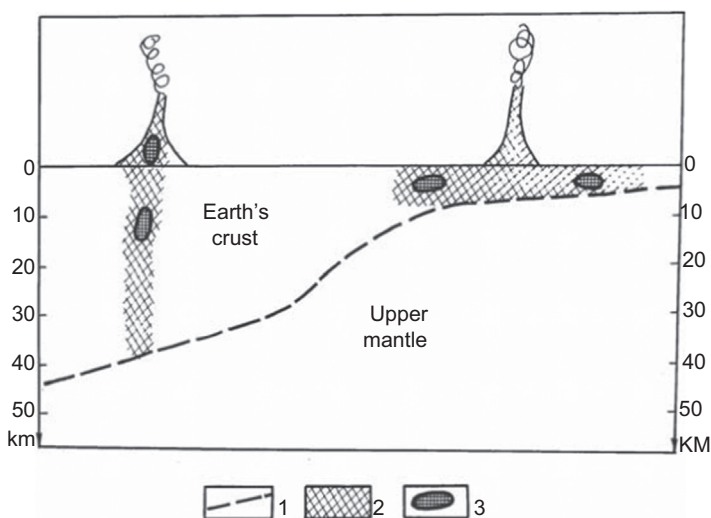


**FIGURE 10.17** Illustration of the difference in the position of the seismic volumes beneath the volcanoes situated above the continental (left) and oceanic (right). Earth's crust (1) "Crust–mantle" boundary, (2) active seismic volume beneath the volcano, and (3) occurrence of the largest volcanic earthquakes. From Zobin, V.M., 1996. *The epicentral areas of volcanic earthquakes and their volcanotectonic significance. Geofísica Internacional* 35, 95–103.

amount of erupted material, there would be a narrow vertical faulted volume beneath the continental volcano and a more widespread and more horizontal faulted volume for the oceanic volcano. Therefore using Eq. (10.3), it is necessary to know the possible deviations of the earthquake epicenter from the volcanic crater. For this purpose, the best-estimated locations of 11 significant volcanic earthquakes presented in Table 10.1 were used (Fig. 10.18). The following equation was obtained (Zobin, 2001):

$$\log D(\text{km}) = 3.67 - 2.14 \log H(\text{km}) \quad (10.4)$$

Eq. (10.4) allows the correct calculation of possible seismic intensity from Eq. (10.3) by moving the position of the initial point from the crater to distance  $D$ , which is the characteristic of the volcano under study. For thick oceanic crust, we have a larger distance between the crater and earthquake epicenter than for thin continental crust.

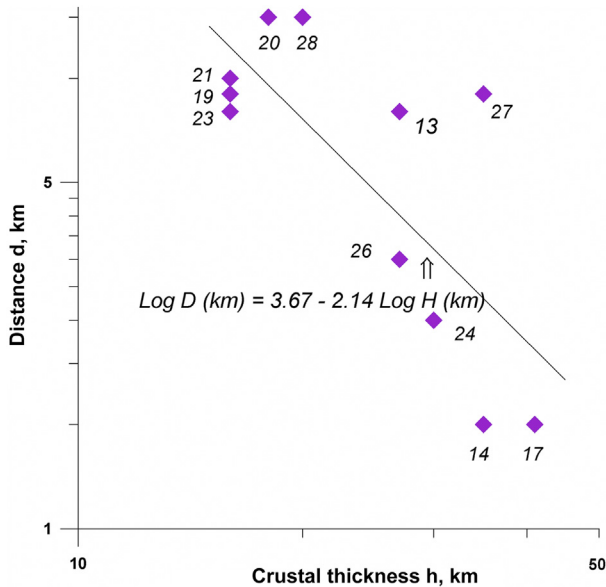


**FIGURE 10.18** Dependence of distance between the crater and the largest earthquake epicenter on the crust thickness in the volcanic region. The line of regression is shown. The coefficient of correlation is  $-0.746$ . The numbers near the black diamonds correspond to the numeration of eruptions in Table 10.1. The position of epicenters is shown in Fig. 10.2. The data about the Earth's crust thickness for events Nos. 13, 14, 17, 19, 20, 21, 23, 26, and 27 are collected in Zobin, V.M., 1996. The epicentral areas of volcanic earthquakes and their volcanotectonic significance. *Geofísica Internacional* 35, 95–103; for the remaining events Nos. 24 and 28, data are taken, respectively, from Mooney, W.D., Laske, G., Masters, T.G., 1998. CRUST 5.1: a global crustal model at  $5^\circ \times 5^\circ$ . *Journal of Geophysical Research* 103, 727–747 and Staples, R.K., White, R.S., Brandsdottir, B., Menke, W., Maguire, P.K.H., McBride, J.H., 1997. Foe-Iceland ridge experiment. 1. Crustal structure of northeastern Iceland. *Journal of Geophysical Research* 102, 7849–7866.

The law of decreasing of earthquake intensity with distance for significant volcanic earthquakes may serve for the prediction of a possible seismic hazard of eruptive activity. This law was calculated based on data from rather deep-seated foci of volcanic earthquakes (focal depth 3–15 km), and therefore it could predict maximum intensity. More shallow seismic sources situated within the volcanic edifice would produce earthquakes of higher attenuation and their intensity would decrease more rapidly.

### 10.4.3 Recurrence Time

Recurrence time  $T$  is the time between similar types of eruptions of the same volcano that occurred at the same location. The data for the estimation of  $T$  were taken mainly from the catalog of [Simkin and Siebert \(1994\)](#). As was shown earlier, the volcano-tectonic activity could be different for different slopes of a volcano ([Gresta et al., 1990](#)). Therefore for the 1914 eruption that occurred on the western and eastern flanks of Sakurajima (Event No. 3), it was taken at a time  $T = \infty$  but not  $T = 135$  years, counting from the 1779 eruption that occurred on the northern and southern slopes of the volcano.

For the 1962 Miyakejima eruption (Event No. 6) that occurred on the northeastern flank with the formation of a radial fissure, the recurrence time was taken  $T = \infty$  because the previous fissure-flank eruptions had all occurred on the other flanks (the 1835 eruption occurred on the western flank and the 1643 eruption occurred on the southwestern flank). At the same time, the 1983 eruption of Miyakejima had a recurrence time  $T = 271$  years because the similar 1712 eruption occurred on the same southwestern flank. The 2000 submarine flank eruption (Event No. 30) occurred on the western flank and has a recurrence time  $T = 165$ . The information about the magnitude of earthquakes associated with the ancient eruptions is absent; therefore, the seismic aspect of the eruption was not taken into account for the determination of  $T$ .

The majority of information about volcanoes is collected for historical time, which is different for different parts of the world. Nevertheless, it is possible to hope that for the volcanoes of study, listed in [Table 10.2](#), we have consistent information during the historical time for the last 100–150 years. For our 32 eruptions with significant seismic events, the value of  $T = \infty$  (the first event for historical time) is available for 23 events out of 32. What does it mean practically? Three volcanoes (No. 4, Paricutin; No. 14, Tolbachik; and No. 32, Dabbahu) were newborn volcanoes. Two submarine eruptions (No. 11, Curacoa Reef and No. 23, Teisi Knoll) were also newborn volcanoes. Several volcanoes (No. 2, Katmai; No. 5, Bezymianny; No. 9, Mount St. Helens; No. 24, Pinatubo; and No. 27, Akademia Nauk) were considered to be dormant volcanoes. For 12 volcanoes, the  $T = \infty$  means that it was the first recorded type of such activity in historical times. For example, for the Usu volcano, the flank eruption of 1910 (No. 1) was the first

mentioned flank eruption after a series of central eruptions beginning in 1663. For Oshima volcano, the eruption of 1986 (No. 21) was the first historical eruption with the forming of a fissure on the north flank after a series of central eruptions of 1783–1974 and previous central and flank eruptions.

The remaining nine eruptions, with  $T$  values less than  $\infty$ , have these values in the range between 110 and 271 years (5 events) and between 12 and 58 years (4 events).

The association between the large recurrence periods for eruptions of the same type at the volcano with significant earthquakes is a very important factor for constraining seismic hazards. Our results show that quasi-dormant volcanoes situated within active volcano-tectonic structures and volcanoes with long-term central crater activity represent higher seismic hazards. They may need to form new passageways for magma ascent. At the same time, the practical application of the recurrence time is problematic. We have the real estimates of recurrence time (not  $\infty$ , or less than 200 years) only for six volcanoes (Sheveluch, Usu, Una Una, Beerenberg, Etna, and Grimsvøtn).

#### 10.4.4 Estimation of the Seismic Hazard of Volcanic Activity of Volcán de Colima, Mexico

Volcán de Colima volcano is the most active one in Mexico. The history of its volcanic activity (Simkin and Siebert, 1994) shows that it is characterized by central or central + flank eruptions. The last central + flank eruption occurred in 1869–75, or more than 140 years ago. We have no information about any significant earthquake related with the eruption and cannot estimate the recurrence time of the event. The  $M_{\max}$  for a possible future significant volcanic earthquake could be taken as  $M_w = 5.9$  for the central + flank eruptions.

The eruption of the second type, the central one, began in 1998 and continued till 2016 without any significant volcano-tectonic earthquake.  $M_{\max}$  for possible significant earthquakes associated with central eruptions of Colima could be taken as  $M_w = 5.4$ .

The thickness of the Earth's crust in the Colima volcanic region is about 36 km (Pacheco et al., 1997). Therefore the position of the seismic event from the central crater, according to Eq. (10.4), would be about 2 km, or insignificantly far from the crater. We can propose a focal depth of 5 km.

In the case of central + flank eruption, the  $M_{\max}$  seismic event could produce, according to the macroseismic Eq. (10.3), a V–VI MM intensity (V–VI means that the intensity MM would be 5.5 grades) in Colima city, situated at 32 km from the volcano, VI MM intensity in the village San Marcos, situated at 14 km from the volcano, and VI–VII MM intensity in the village of La Yerbabuena, situated at 8 km from the volcano.

In the case of a central eruption, the intensities MM would be V–VI in Colima city, VI in San Marcos, and in La Yerbabuena.

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## RELEVANT WEBSITES

- BGVN. Bull. Glob. Volc. Network. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.
- Harvard CMT. <http://www.globalcmt.org/CMTsearch.html>.
- ISC, International Seismological Centre. <http://www.isc.ac.uk>.
- JMA, Japan Meteorological Survey. <https://www.jma.go.jp/jma/indexe.html>.

## Chapter 11

# Origin of Eruption Earthquakes

Eruption earthquakes are related to the movement of magma within volcanic conduits and to subsurface or surface manifestations of volcanic activity. They may be produced by magmatic, phreatic, and phreatomagmatic processes within the conduit or the crater of a volcano as well as by volcanic explosions and the propagation of density currents along the slopes of the edifice (lava flows, pyroclastic flows, and lahars). Therefore eruption earthquakes reflect the variety of volcanic processes and understanding them is crucial for effective eruption monitoring and the quantification of the associated phenomena.

### 11.1 VOLCANIC PROCESSES GENERATING SEISMIC SIGNALS OF ERUPTION EARTHQUAKES

During its ascent within the volcanic conduit, magma undergoes exsolution and fragmentation, thus, changing from a continuous melt with a dispersed gas phase to a two-phase mixture of melt fragments within a continuous gas phase (Cashman et al., 2000). This process may lead to effusive and/or explosive eruptions depending on bubble number densities and the magma flow rate (Gonnermann and Manga, 2007). All these processes represent potential sources of seismic signals.

#### 11.1.1 Processes within the Volcanic Conduit

The transport of magma from the deep reservoir to the surface within the conduit can produce different types of seismic signals. Networks of cracks can be formed within the magma body. The cracks, filled by fluid phases, can be excited into resonance by a pressure transient (Chouet, 2003) and generate long-period seismic signals. The upward movement of the fluid—particle mixture above the fragmentation surface can trigger vibrations in the conduit structures (Ripepe et al., 2001). Explosions near or within the crater are possible sources of seismic signals, as well. The fracturing of lava dome rock can generate seismic signals also (Tuffen et al., 2008; see Chapter 3.).

#### 11.1.2 Volcanic Flows

The subaerial massive sediment motions at volcanoes are divided into two groups: the phenomena in which the particles are dispersed in the flowing

body and the phenomena in which the moving bodies are mostly the agglomerates of soil and rocks (Takahashi, 2007). The pyroclastic flows are included in the first group, while lahars belong to the second group.

When magma emerges from the conduit, it may form lava domes and/or lava flows. The type of lava construction depends on the nature of the feeding magma. Basaltic volcanoes form rather fluid low-viscosity lava flows. Andesitic and dacitic volcanoes form lava domes within the crater and/or high-viscosity lava flows. The gravitational collapse (Merapi type) or the explosive disruption of a growing lava dome (Peléean type) and the collapse of an eruption column (Soufrière type) can generate pyroclastic flows and rockfalls (Francis, 1998). Pyroclastic flows are distinguished from rockfalls by their larger size, longer runout ( $\geq 0.5$  km), greater production of fine ash, and appreciable buoyant hot ash clouds (Calder et al., 2002).

The model of Takahashi and Tsujimoto (2000) illustrates that pyroclastic flow propagation can be represented as a two-stage process: the collapse of a lava dome (or the rocks from the eruptive column) induces a granular flow, which later changes to a fluidized flow (Fig. 11.1A). As was shown by Yamasato (1997) and Zobin et al. (2009b), the main part of the seismic signal is associated with the fluidized flow stage of pyroclastic flows.

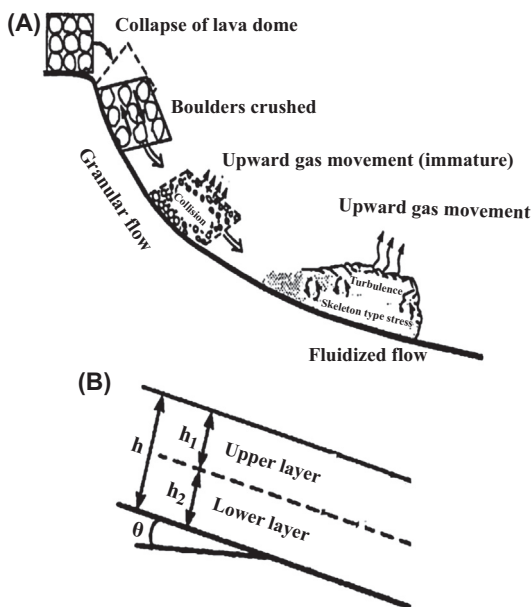


**FIGURE 11.1** Mechanical models of pyroclastic flows (A) and lahars (B). Details are discussed in text. After Takahashi, T., Tsujimoto, H., 2000. A mechanical model for Merapi-type pyroclastic flow. *Journal of Volcanology and Geothermal Research* 98, 91–115 and Takahashi, T., Satofuka, Y., 2002. Generalized theory of stony and turbulent muddy debris flow and its practical model. *Journal of the Japan Society of Erosion Control Engineering* 55, 33–42.



Lahars along the slopes of volcanoes are usually triggered by heavy rain, especially after the intensive fall of fresh ash. At Merapi volcano, for example, about 20 mm/h is supposed to be critical for lahar triggering if it falls on unconsolidated materials such as ash, sand, or gravel (Lavigne et al., 2000). In some instances, lahars may have been triggered by steam explosions that generated strong ground vibrations, which induced the failure of partially water-saturated pyroclastic materials (Rodolfo et al., 1996). The typical 1991 posteruption lahars at Pinatubo were estimated to be 2–3 m deep, 20–50 m wide, and moving at 4–8 m/s (Pierson et al., 1996). The frontal mean velocity of lahars at Merapi ranges between 2.5 and 5.9 m/s; the peak flow velocity may reach 6–15 m/s (Lavigne et al., 2000). Small- to moderate-sized lahars at Pinatubo remobilized between  $2 \times 10^5$  and  $10^6$  m<sup>3</sup> of sediment; larger events lasting up to 4 hours have transported about 10 times this amount (Rodolfo et al., 1996).

Takahashi and Satofuka (2002) describe lahars as hybrid-type debris flows between turbulent-muddy and stony debris flows. The hybrid flow in the inertial range consists of two layers: the upper layer of depth  $h_1$  is the particle suspension layer and the lower layer of depth  $h_2$  is where particles move with frequent collision or enduring contact with each other (Fig. 11.1B). The dominance of stony or turbulent-muddy type of flow depends on the relative thickness of these two layers  $h_1/h_2$ . The movement of the lower stony layer along the volcanic slope ravine (angle of inclination  $\theta$  in Fig. 11.1B) produces the main part of the seismic signal of lahar.

## 11.2 SOURCE MECHANISMS OF ERUPTION EARTHQUAKES

Source mechanisms of eruption earthquakes may be represented by a combination of single forces and the seismic moment tensor as was described in Chapter 3. In this chapter, we describe the applications of these characteristics of the source mechanisms for the special cases of eruption earthquakes.

### 11.2.1 A Force System Equivalent to a Volcanic Eruption

The model (Kanamori and Given, 1982, 1983; Kanamori et al., 1984) is based on the assumption that seismic signals are produced by a volcanic eruption as Lamb pulses, excited by a nearly vertical single force that represents the counter force of the eruption. Lamb (1904) computed the transient response of a homogeneous elastic half space to a single force with various geometries. He showed that a pronounced pulse propagates along the free surface with the Raleigh-wave velocity. This pulse is preceded by minor signals, which propagate with P- and S-wave velocities (Fig. 11.2).

The comparison of observed seismograms for small eruptions at Mount St. Helens with synthetics constructed as Lamb's pulses excited by a vertical force (Fig. 11.3) shows a good agreement for the Rayleigh-wave signal. Forward modeling allows estimating the counter force of the eruption by

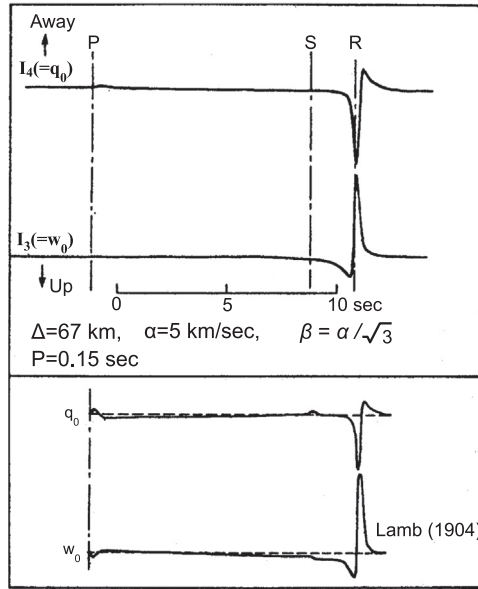


**FIGURE 11.2** Lamb pulses computed for a downward vertical single force applied at the free surface of a homogeneous half-space at a distance of 67 km.  $\alpha$  and  $\beta$  are P- and S-wave velocities,  $I_3$  and  $I_4$  correspond to vertical and radial components of displacement, respectively.  $I_3$  and  $I_4$  are convolved with the time derivative of a source-time function of the form  $s(t) = p/(p^2 + t^2)$  as proposed by Lamb (1904). Corresponding traces given by Lamb (1904) are also shown for comparison.  $I_3$  and  $I_4$  correspond to  $w_0$  and  $q_0$ . From Kanamori, H., Given, J.W., 1983. Lamb pulse observed in nature. *Geophysical Research Letters* 10, 373–376.

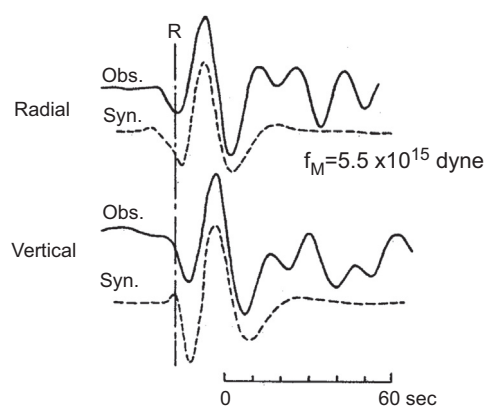


**FIGURE 11.3** Comparison of the synthetic for the vertical force with the observed long-period record of the June 13, 1980 Mount St. Helens explosion at a distance of 67 km from the volcano. The synthetics are computed using a distance of 67 km,  $\mu = 1.9 \times 10^{11}$  dyne/cm<sup>2</sup>, and  $f_M$  (peak value of the force) =  $5.6 \times 10^{15}$  dyne. From Kanamori, H., Given, J.W., 1983. Lamb pulse observed in nature. *Geophysical Research Letters* 10, 373–376.

comparing the amplitude of the synthetic with the recorded signal. Modeling of more complex signals can be performed assuming that a volcanic eruption may be considered as a combination of a single force, representing the thrust of the eruption, and an isotropic source corresponding to a sudden pressure release in the magma chamber (Fig. 11.4). For a simple model of eruption, the amplitude ratio of seismic waves excited by a single force to those excited by an isotropic source is approximately equal to the ratio of the seismic wave velocity to the particle velocity of the fluid or gas in the magma chamber. Since this ratio is about 10, a single force can adequately model the source (Kanamori and Given, 1983).

Modeling of the complex seismic signal produced by the great May 18, 1980 Mount St. Helens eruption is shown in Fig. 11.5. The eruption consisted of separate events that contributed to shaping the overall seismic signal. A chronology of the eruptive sequence is shown in Fig. 11.5B. The sequence began with a magnitude  $M_s$  5.2 earthquake. Within 7–20 seconds of the earthquake, a 2.3-km<sup>3</sup> landslide began (Event 1). During the landslide, an explosion (the lateral blast) occurred (Event 2). A second explosion occurred just after the termination of the landslide. Two minutes after the first earthquake, a second earthquake occurred. The sequence lasted about 3 minutes.

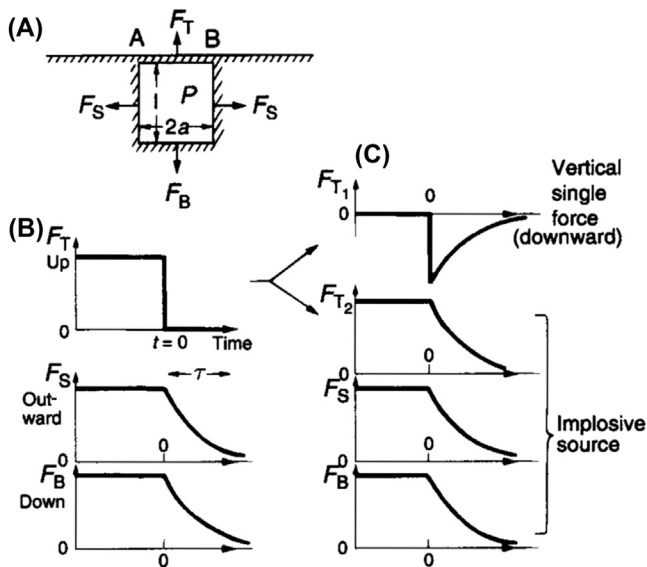


**FIGURE 11.4** Force system equivalent to a volcanic eruption. (A) A pressure release model for a volcanic eruption. (B) Forces acting on the top, side, and bottom walls. (C) Decomposition of the force into a vertical single force and an implosive force. From Kanamori, H., Given, J.W., Lay, T., 1984. Analysis of seismic body waves excited by the Mount St. Helens eruption of May 18, 1980. *Journal of Geophysical Research* 89, 1856–1866.

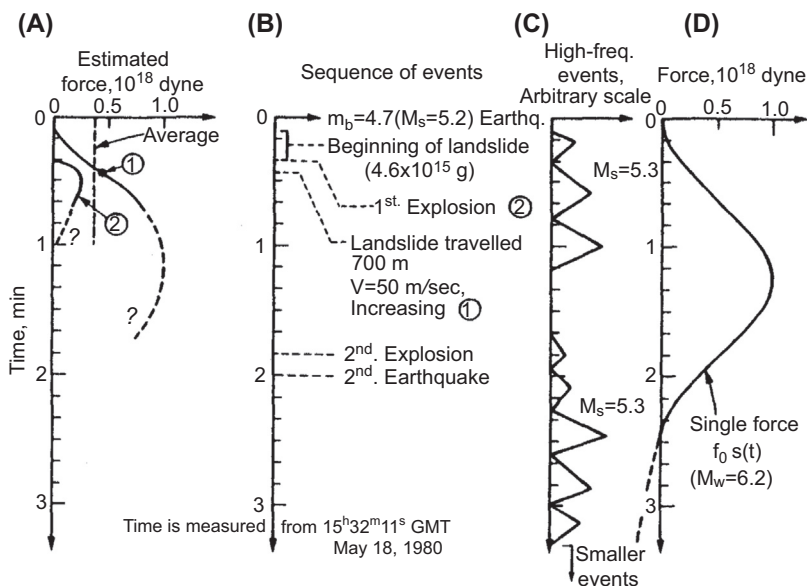


**FIGURE 11.5** Schematic diagrams showing the chronology of events associated with the Mount St. Helens eruption of May 18, 1980. Time is taken positive downwards. The time history of the single force determined from surface waves is shown in the right panel (A and B), and the magnitude and the time history estimated from the reported sequence are shown in the left panel (C and D). From Kanamori, H., Given, J.W., 1982. Analysis of long-period seismic waves excited by the May 18, 1980, eruption of Mount St. Helens—a terrestrial monopole? *Journal of Geophysical Research* 87, 5422–5432.

The time history of the single force inferred from the long-period waves (Fig. 11.5D) characterizes the entire eruption sequence as a single event equivalent to an earthquake of magnitude  $M_w$  6.2. The value of this force reached its maximum when the landslide had a maximum velocity. The source-time function inferred from the P-wave deconvolved record (Fig. 11.5C) allows interpretation of the sequence of events. There are two separate events of the same magnitude (5.3) that might be associated with the first earthquake + first explosion and the second earthquake + second explosion (Fig. 11.5B).

Fig. 11.5A shows the comparative modeling of the time history of the eruption with the magnitude of the forces caused by the landslide and the lateral blast. According to this reconstruction, the most energetic process was the landslide; the resulting force peaked at about  $10^{18}$  dyne. The lateral blast was less intense; its resulting force reached a maximum of  $3.4 \times 10^{17}$  dyne.

### 11.2.2 Seismic Moment Tensors of Some Non–Double-Couple Sources of Eruption Earthquakes

Seismic moment tensors describing volumetric mechanisms are thought to represent magma-filled sources of eruption earthquakes. While the injection

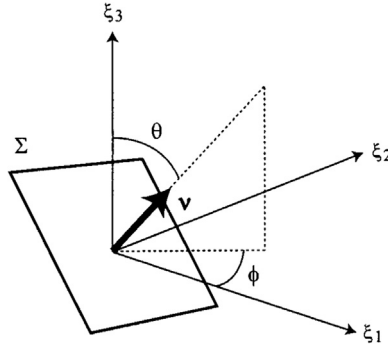


**FIGURE 11.6** System of coordinates and geometry for a crack.  $\xi_1$ ,  $\xi_2$ , and  $\xi_3$  correspond to the east, north, and vertical directions, respectively;  $\lambda$  and  $\mu$  are the elastic parameters of the host rocks. From Kumagai, H., 2009. *Volcano seismic signals, source quantification of*. In: Meyers, R.A. (Ed.) *Encycl. Complexity and System Sci.*, Springer-Verlag, Berlin, pp. 9899–9932.

of a volcanic fluid or a thermal expansion of a fluid-filled crack occurs, a sudden volume increase  $\Delta V$  is observed and the fracture acts as a radiating seismic source. The seismic moment tensors of these sources can be represented in a Cartesian system of coordinates (Fig. 11.6).

#### 11.2.2.1 Tensile Crack Source

Source coordinates and geometry for the tensile crack are shown in Fig. 11.6 (Kumagai, 2009). Here  $v$  is the direction of opening-closing described by the angles  $\theta$  and  $\varphi$ , where  $\theta$  is the angle of measured from the up direction  $\xi_3$  and  $\varphi$  is the angle of horizontal projection of measured from direction  $\xi_1$ . The moment tensor for a vertical tensile crack ( $\theta = \pi/2$ ) undergoing a volume change  $\Delta V$  is given as follows:

$$\vec{M} = \mu \Delta V \begin{pmatrix} \frac{\lambda}{\mu} + 2\cos^2\varphi & 2\sin\varphi\cos\varphi & 0 \\ 2\sin\varphi\cos\varphi & \frac{\lambda}{\mu} + 2\sin^2\varphi & 0 \\ 0 & 0 & \frac{\lambda}{\mu} \end{pmatrix} \quad (11.1)$$

and for a horizontal crack is

$$\vec{M} = \mu \Delta V \begin{pmatrix} \frac{\lambda}{\mu} & 0 & 0 \\ 0 & \frac{\lambda}{\mu} & 0 \\ 0 & 0 & \frac{\lambda}{\mu} + 2 \end{pmatrix} \quad (11.2)$$

If  $\lambda = \mu$  and  $\varphi = 0$ , the amplitude ratio of the diagonal terms is 3:1:1 for a vertical crack and 1:1:3 for a horizontal crack.

### 11.2.2.2 Cylindrical Source

The surface of a cylinder can be regarded as a cylindrical crack (Kumagai, 2009), where radial expansion,  $D_c = (d_c + \Delta_c)$ , occurs (Fig. 11.7). The moment tensor for the radial expansion of a vertical cylinder of length  $L$  and radius  $R$  is given by

$$\vec{M} = \Delta V \begin{pmatrix} \lambda + \mu & 0 & 0 \\ 0 & \lambda + \mu & 0 \\ 0 & 0 & \lambda \end{pmatrix} \quad (11.3)$$

where  $\Delta V$  is the volume change given as

$$\Delta V = 2\pi R L D_c = 2\pi R L (d_c + \Delta_c) \quad (11.4)$$

If  $\lambda = \mu$ , the amplitude ratio of the diagonal terms is 2:2:1.

### 11.2.2.3 Spherical Source

The moment tensor for a spherical source (Fig. 11.8) is given as (Kumagai, 2009)

$$M = (\lambda + 2\mu) \Delta V_s \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix} \quad (11.5)$$

where  $\Delta V_s = 4\pi R^2 \Delta_s$  is the volume change caused by the displacement  $\Delta_s$  of a spherical surface of radius  $R$ . The amplitude ratio of the diagonal terms is 1:1:1.

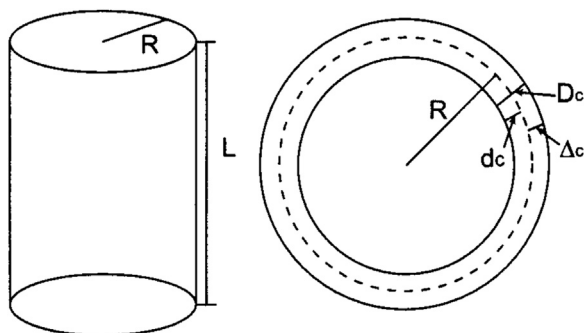


**FIGURE 11.7** Vertical cylinder of length  $L$  and radius  $R$ . The cylinder surface can be regarded as a cylindrical crack, where radial expansion  $D_c = (d_c + \Delta_c)$  occurs. The inner wall of the crack moves inward by  $d_c$  and the outer wall moves outward by  $\Delta_c$ . From Kumagai, H., 2009. *Volcano seismic signals, source quantification of*. In: Meyers, R.A. (Ed.) *Encycl. Complexity and System Sci.*, Springer-Verlag, Berlin, pp. 9899–9932.

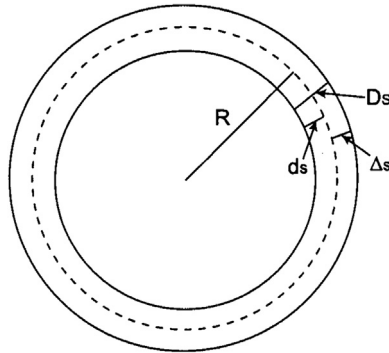


**FIGURE 11.8** Spherical crack surface of radius  $R$  where a constant radial expansion  $Ds = (ds + \Delta s)$  occurs. The inner wall of the crack moves inward by  $ds$  and the outer wall moves outward by  $\Delta s$ . From Kumagai, H., 2009. *Volcano seismic signals, source quantification of*. In: Meyers, R.A. (Ed.) *Encycl. Complexity and System Sci.*, Springer-Verlag, Berlin, pp. 9899–9932.

## 11.3 MODELS OF THE ERUPTION EARTHQUAKE SOURCES

### 11.3.1 Models Based on the Vibration of Magma-Filled Structures

The models have been developed in two main directions: (1) those based on the resonance of fluid-filled cavities with various geometries (sphere, pipe, or crack) and (2) those based on the oscillation of a number of vibrating conduits. A seismic signal is generated by pressure fluctuations in the magma system. The choice of the model is mostly subjective and depends on the knowledge of specific characteristics of the volcano, as well as on the experience of scientists.

#### 11.3.1.1 The Fluid-Driven Crack Model

The model (Chouet, 1986) consists of a rectangular fluid-filled crack contained within an infinite, linear elastic medium (Fig. 11.9). The thickness of the crack is much smaller than the seismic wavelengths of interest. The fluid-driven crack is a composite of a trigger element (the energy source) and a resonator (the fluid-filled crack). A pressure transient is applied over a small area of the crack wall triggering the acoustic resonance of the crack. Repetitive triggering of resonance in the crack by fluid flow produces continuous tremors.

The resonance characteristics of the fluid-filled crack depend on the crack geometry, the physical properties of the fluid and solid, the impedance contrast between fluid and solid, the spatiotemporal characteristics of the pressure fluctuations driving the crack, and the boundary conditions in effect at the crack perimeter. The impedance contrast affects both the frequency and duration of the signal radiated by the crack.

The ground response to a buried source may be synthesized after calculating the source space-time function. Variations in the model characteristics enable the optimization of the synthetic modeling. Fig. 11.10 shows the synthetics

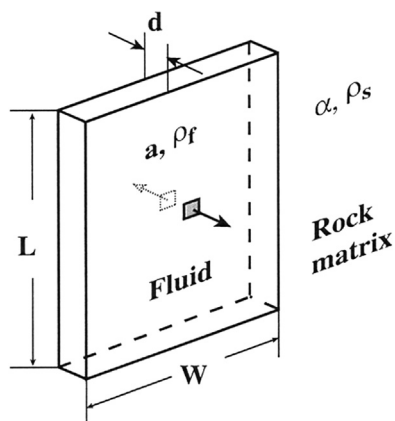


**FIGURE 11.9** Geometry of the fluid-filled crack model. From Chouet, B., 2003. *Volcano seismology. Pure and Applied Geophysics* 160, 739–788.

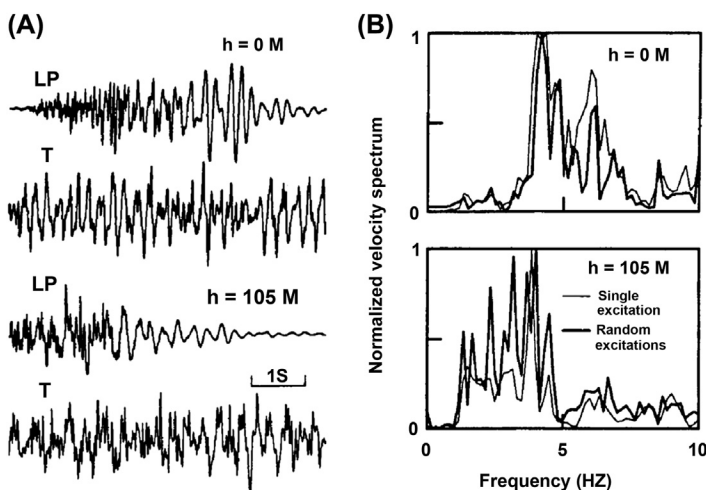


**FIGURE 11.10** Dependence of synthetics on crack depth and crack excitation. All synthetics are for a 100-m-wide, 200-m-long, vertical crack excited by a pressure transient applied at the center of the crack. Each synthetic represents a vertical component of ground velocity calculated at a distance of 1 km and azimuth of 25 degrees. (A) The synthetics of long-period (LP) events and volcanic tremor (T) calculated for the crack depths of 0 m and 105 m. (B) The Fourier spectra of the same records. From Chouet, B., 1996b. *Long-period volcano seismicity: its source and use in eruption forecasting. Nature* 380, 309–316.

calculated for different crack excitations and crack depths. The synthetic shape strongly depends on the model depth (compare the synthetics calculated for the depths of 0 m and 105 m). Fig. 11.10 demonstrates that the single excitation synthetics give a long-period event waveform, while the synthetics obtained by delaying and superposing multiple copies of the same signal, give the tremor record. Fig. 11.11 illustrates a good fit of the synthetics to the observed record of a long-period event recorded at Galeras volcano, Colombia.



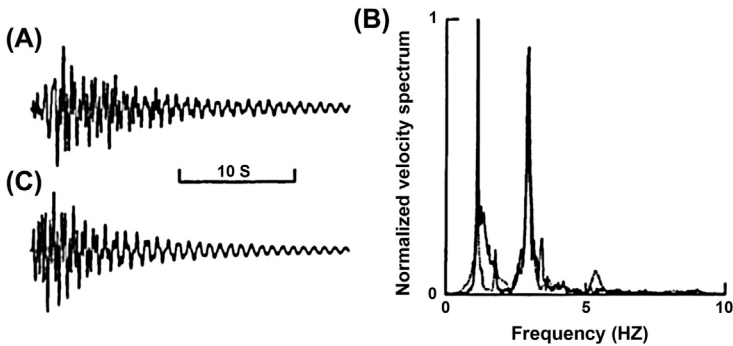


**FIGURE 11.11** Comparison between the vertical ground velocity recorded at a distance of 1 km from the crater (A) and the synthetics calculated at the same location for the excitation of a vertical fluid-filled crack buried at a depth of 1 km beneath the vent of a volcano (B) for the long-period event that preceded the January 14, 1993 eruption of Galeras volcano, Colombia. The normalized spectra for data (thin line) and synthetics (dotted line) are shown in (C). From Chouet, B., 1996b. Long-period volcano seismicity: its source and use in eruption forecasting. *Nature* 380, 309–316.

#### 11.3.1.2 The Conduit-Vibration Model

This model (Seidl et al., 1981; Schick et al., 1982; Gresta et al., 1991) may be applied to volcanic tremor. It is hypothesized that transient flows within volcanic conduits can generate tremors with certain dominant frequencies, depending on the conduit geometry. Vapor and gas escaping through open conduits are considered responsible for turbulent motions in the magma; consequently, the magma-filled pipes in the upper structures of the volcano behave as oscillators. Fig. 11.12 shows a model of conduit oscillators for Etna volcano, Sicily. It consists of a set of vibrators with characteristic frequencies from 0.9 to 4 Hz. These characteristic frequencies are comparable with the peak frequencies that were observed for volcanic tremor at Etna (Fig. 11.13).

### 11.3.2 Models Based on the Process of Deep Ascending of Magma Before an Explosion

The following two models consider the deep trigger of an explosion associated with the vertical magma ascent from the conduit bottom to the surface.

#### 11.3.2.1 Two-Stage Model of Volcanic Explosion

The contemporary records of broadband seismic signal and video images of a Vulcanian explosion at Volcán de Colima (Zobin et al., 2009b) showed that the first superficial manifestation of the explosion (outlet of magma material from the crater) appears only about 15 seconds after the onset of the seismic record, on the code of the second pulse (Fig. 11.14). This indicates the ascending process occurring within the volcano conduit before an explosion. The model proposed by Zobin et al. (2006a,b, 2009a) describes the volcanic explosion process as consisting of the

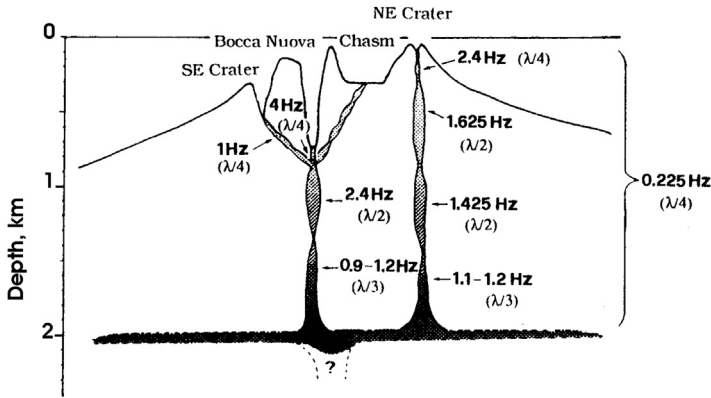


**FIGURE 11.12** Schematic model of main magma conduits at Etna. Bocca Nuova and Chasm are active craters of Etna. From Schick, R., Cosentino, M., Lombardo, G., Patané, G., 1982a. Volcanic tremor at Mount Etna. A brief description. *Memorie—Società Geologica Italiana* 23, 191–196.

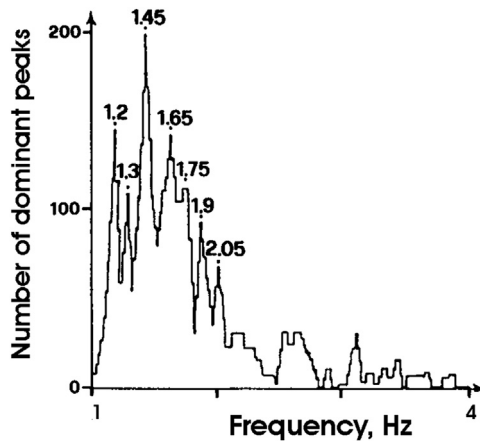


**FIGURE 11.13** Histogram of the first three dominant peaks of tremor recorded from January 1973 to January 1985 at Etna. From Cosentino, M., Lombardo, G., 1989. A model for internal dynamical processes on Mt. Etna. *Geophysical Journal International* 97, 367–379.

ascent of fragmented magma to the surface, triggering an explosion (stage 1), and a subsequent explosion (stage 2; Fig. 11.15A). Based on this model, the seismic record of an explosion (Fig. 11.15B) may be considered as consisting of two phases, preexplosion (between  $t_1$  and  $t_2$ ) and coexplosion (between  $t_2$  and  $t_3$ ).

This model may be applied for quantification of the sources of Strombolian and Vulcanian explosions. The initial seismic pulses, generated by a Strombolian explosion at the basaltic volcano, are supposed to be produced by the rapid expansion of gas in the magma conduit (Ripepe et al., 2001). For more viscous magmas, typical for Vulcanian explosions at

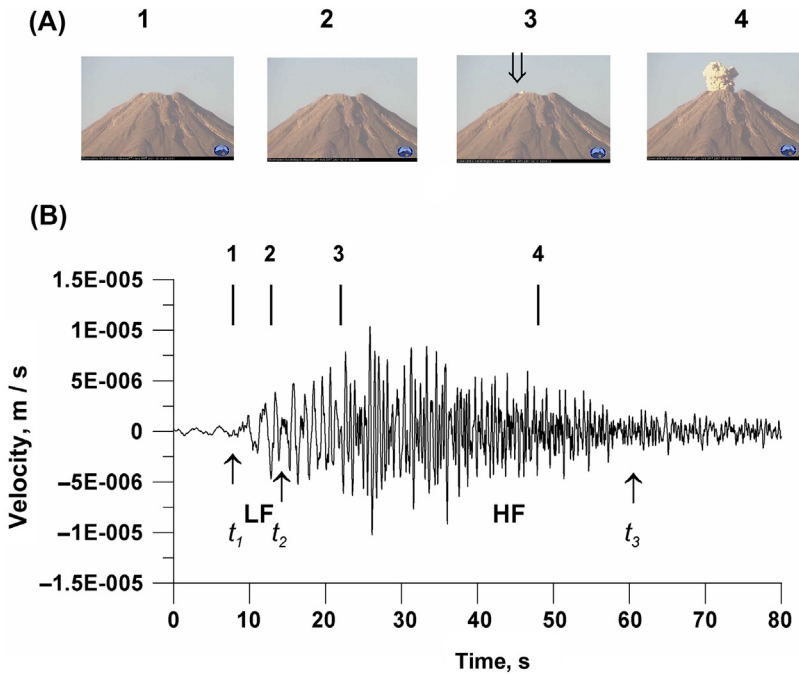


**FIGURE 11.14** (A) Comparison of the broadband seismic record (EZ5 station) of an explosion at Volcán de Colima (February 21, 2007) with the video images of the same event. (B) 1–4 corresponds to the timing of images and seismic record;  $t_1$ ,  $t_2$ , and  $t_3$  are the times of the onset and the end of preexplosion (P\_E) pulse, and the end of coexplosion (C\_E) pulse, respectively; the arrow shows the start of ejection of explosive material from the crater. The digital infrared video camera was installed at a distance of 15 km from the crater and records a frame/second. The timing at seismic station and for the video camera was synchronized by the world atomic time survey. From Zobin, V.M., Reyes, G.A., Guevara, E., Bretón, M., 2009b. Scaling relationship for Vulcanian explosions derived from broadband seismic signals. *Journal of Geophysical Research* 114, B03203.

andesitic and dacitic volcanoes, the initial seismic pulse is supposed to be produced by gas filtration through a system of interconnected bubbles (Melnik and Sparks, 1999). The coexplosion pulses are supposed to be produced by an explosion for both Strombolian and Vulcanian activity.

The four source parameters of an explosion may be inferred from broadband seismic records: the force,  $F$ , acting in the conduit before an explosion and governing the ascent of fragmented magma to the surface, the time,  $D_1 = t_2 - t_1$ , corresponding to the migration of fragmented magma in the conduit before an explosion, the duration of explosion in the conduit,  $D_2 = t_3 - t_2$ , and the energy,  $E$ , of the explosion. The power of an explosion  $P$  may be obtained as  $E/D_2$ .

The Rayleigh nature of seismic waves representing the preexplosion seismic pulses shown by Zobin et al. (2006a,b) allows using the

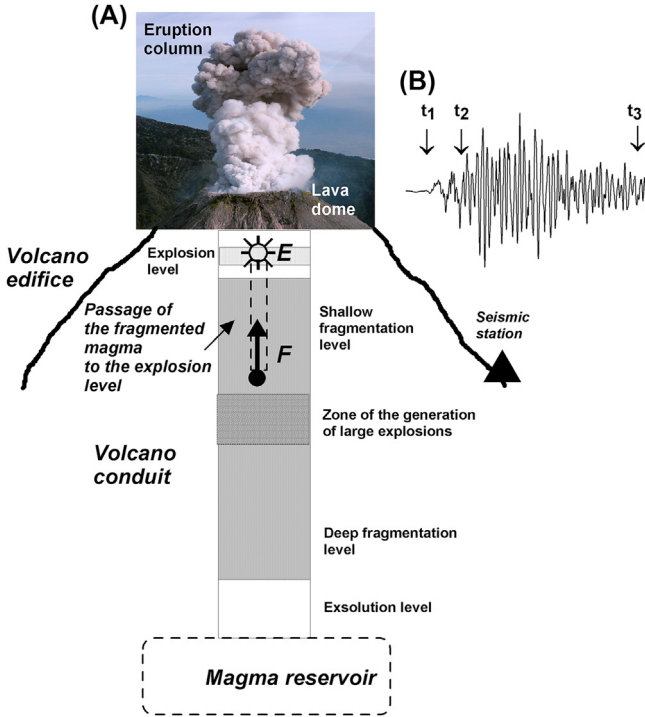


**FIGURE 11.15** Illustration of the two-stage model of a Vulcanian explosion (A) and its broad-band seismic record (B). Photo of the Volcán de Colima eruption column (December 2004) was taken by C. Navarro. From Zobin, V.M., Reyes, G.A., Guevara, E., Bretón, M., 2009b. Scaling relationship for Vulcanian explosions derived from broadband seismic signals. *Journal of Geophysical Research* 114, B03203.

methodology of modeling the first pulse as a Lamb's pulse with the estimation of the force  $F$  (Kanamori and Given, 1983; Nishimura, 1995; Fig. 11.16A). The energy of the explosion,  $E$ , may be calculated from the spectrum of coexplosion pulse (Fig. 11.16B). Parseval's theorem states (Weaver, 1983) that it is possible to calculate the total amount of energy (for the unit of mass), contained in a continuous signal  $v(t)$ , from the continuous Fourier transform  $V(f)$ . In physics and engineering, Parseval's theorem is often written as follows:

$$\int_{-\infty}^{+\infty} |v(t)|^2 dt = \int_{-\infty}^{+\infty} |V(2\pi f)|^2 df \quad (11.6)$$

The left side of Eq. (11.6) represents what in data processing is called "the total energy of signal  $v(t)$ ." It is assumed that the energy is finite even if the signal may mathematically have an infinite length. The right side then

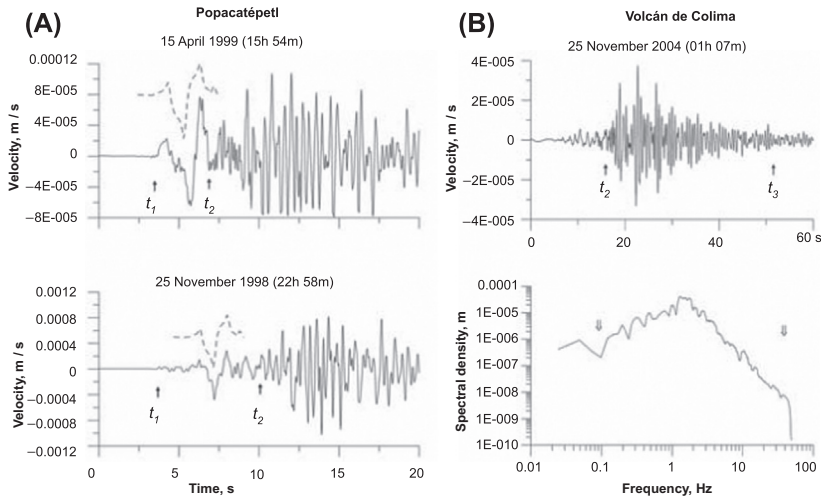


**FIGURE 11.16 Illustration of the processing of seismic signals.** The broadband unfiltered seismic signals (vertical component of velocity) corrected for instrument response are shown. The time indices  $t_1$ ,  $t_2$ , and  $t_3$  are the same as in Fig. 11.14. (A) The modeling of the preexplosion signals for the seismic records of the Popocatepetl explosions. The *dashed line* shows the synthetic seismogram; the *solid line* is the observed seismogram. (B) The seismic record of the Colima explosion and the Fourier spectrum of the coexplosion impulse. The range of frequencies taken for calculation of energy is shown by the *double hollow arrows*.

represents the energy as an integral over all frequencies  $f$ . For the application of this theorem for calculation of the energy of discrete-time waveform  $v(t)$  of the seismic signal of duration  $T = t_i - t_j$ , the range of frequencies in the right integral may be selected as shown in Fig. 11.16B.

To estimate the total energy of the explosion, the energy of the coexplosion impulse,  $E_i$ , is multiplied by the coefficient of the seismic portion of the total energy of the explosion (the ratio between seismic and total energy)  $p$  and the coefficient of effective attenuation of seismic energy with distance including the effects of geometrical spreading, energy absorption, focal mechanism, and station conditions,  $k$ . The coefficients  $p$  and  $k$  were approximately estimated for the Colima explosions recorded by the broadband seismic station at a distance of 4 km from the crater: The coefficient of the seismic portion of the total energy of an explosion  $p$  was estimated equal to  $10^{-5}$  and the coefficient of effective attenuation of seismic energy with distance  $k$  was equal to  $10^{-13}$ . As a result, the following equation for estimation of the energy of explosions at Volcán de Colima at the distance of 4 km from the crater was proposed:

$$E = 10^{18} \times E_i \quad (11.7)$$

Eq. (11.7) provides the values of  $E$  with a precision of about half an order of magnitude. At the same time, the estimations of  $E$  for a sequence of explosions at the same volcano can give a resolution of about  $\pm 0.1$  of the log unit.

### 11.3.2.2 Ascending Seismic Source Model of the Source Process of an Explosion

The model developed by Kumagai et al. (2011) proposes that a sudden pressure drop at the base of the supersaturated magma in the conduit triggers the growth of bubbles and generates a pressure wave that propagates to the surface along the magma conduit with the generation of seismic waves. The exsolution of steam from the magma produces the gas-volume fractions increasing with the depth decrease and resulting in fragmentation of the magma at shallow depths. Fragmented magma with gases explodes at the surface. A pressure disturbance in the magma at depth and its upward propagation are considered as the fundamental processes that trigger an explosive eruption (Kumagai et al., 2011).

This model was proposed based on the source study of the explosion event of February 11, 2010, at the andesitic stratovolcano Tungurahua, Ecuador. Five broadband seismic stations were in operation around the volcano (Fig. 11.17A). Two methods were applied for the location of the ascending seismic sources of

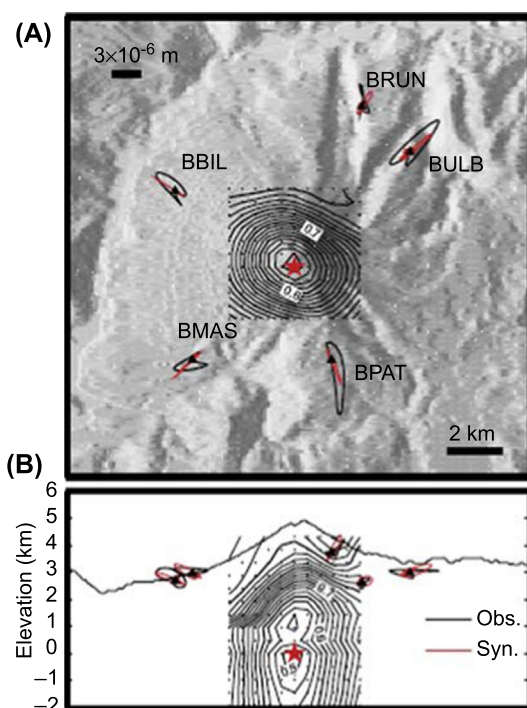


**FIGURE 11.17** Location of the source of the February 11, 2010 explosion at Tungurahua volcano by waveform inversion: (A) horizontal projection and (B) vertical projection. In (A), the broadband seismic stations used in waveform inversion are shown with triangles. Contour plots of horizontal and vertical distributions of the normalized residuals obtained by waveform inversion are shown. Dots in (B) indicate the nodes used in the grid search to estimate the source location. Observed and synthetic particle motions obtained from displacement seismograms are shown in (A) and (B) with black and red lines (Gray in print versions), respectively. Modified from Kumagai, H., Placios, P., Ruiz, M., Yepes, H., Kozono, T., 2011. Ascending seismic source during an explosive eruption at Tungurahua volcano, Ecuador. *Geophysical Research Letters* 38, L01306.

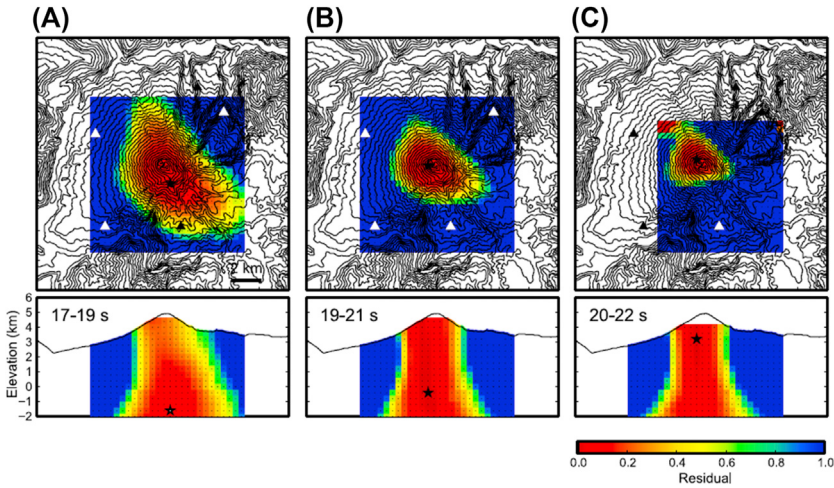


**FIGURE 11.18** A sequence of three locations of the source of the February 11, 2010 explosion at Tungurahua volcano in three time periods (A, 17–19 s; B, 19–21 s; and C, 20–22 s). The horizontal and vertical distributions of the normalized residual estimated for explosion event are shown. Stars indicate the minimum residual points. Dots indicate the nodes used in the grid search to estimate the source location. From Kumagai, H., Placios, P., Ruiz, M., Yepes, H., Kozono, T., 2011. Ascending seismic source during an explosive eruption at Tungurahua volcano, Ecuador. *Geophysical Research Letters* 38, L01306.

this explosive event. Waveform inversion of low-frequency signals at the event onset pointed to the source with an isotropic mechanism at a depth of around 5 km beneath the summit (Fig. 11.17B). A source location method using high-frequency seismic amplitudes with an *S*-wave velocity of 2000 m/s demonstrated that the event was triggered at a depth of around 6 km beneath the summit. The sequence of three source locations (Fig. 11.18) with the residual distributions for three time windows (17–22 seconds) at the event onset indicated that the source ascended toward the summit at a speed of about 1600 m/s.

The source-time function at the best-fit isotropic source location showed initial deflation was followed by inflation (Fig. 11.19). The initial deflation in the source-time function suggests that a sudden pressure drop was the trigger for the explosion (Kumagai et al., 2011).

### 11.3.3 Modeling of Seismic Signals Generated by Pyroclastic Flows and Rockfalls

Pyroclastic flows (or pyroclastic density currents) are the inflated mixtures of hot volcanic particles and gas that flow in variable concentrations and varying velocity along the ground (Schmincke, 2004).

#### 11.3.3.1 Three-Stage Dome Collapse Model

Modeling of the seismic signals produced by pyroclastic flows followed the three elements noted by Takahashi (2007): fall, flow, and slip. The first element is



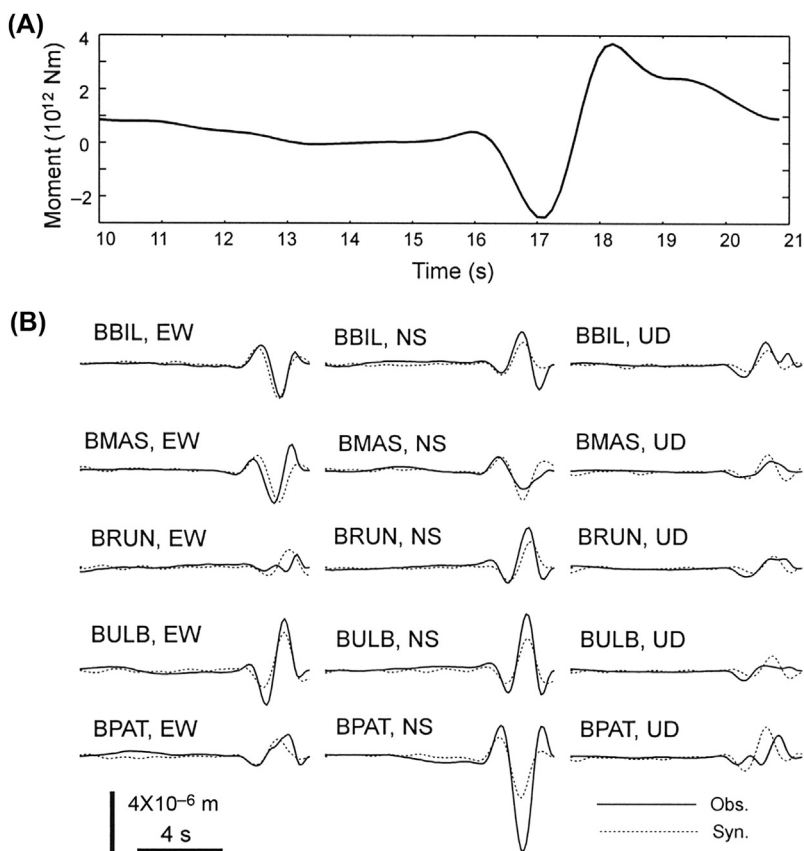


**FIGURE 11.19** The source-time function at the best-fit isotropic source location of the explosive event of February 11, 2010 at Tungurahua volcano obtained by waveform inversion (A). The observed and synthetic waveforms used in inversion are shown in (B). The positions of seismic stations are shown in Fig. 11.17A. From Kumagai, H., Placios, P., Ruiz, M., Yepes, H., Kozono, T., 2011. Ascending seismic source during an explosive eruption at Tungurahua volcano, Ecuador. *Geophysical Research Letters* 38, L01306.

variable; it may be the fall of the massive sediments caused by the collapse of a lava dome segment or the fall of the rocks from a volcanic eruption column. The flow and slip elements would be similar.

The observations of the generation of pyroclastic flows at Unzen volcano, Japan (Yamasato, 1997; Uhira et al., 1994) illustrated the three stages in the formation of a pyroclastic flow with a volume equal to  $5.7 \times 10^4 \text{ m}^3$ , recorded by video and seismic instruments (Fig. 11.20): the lava dome collapse (stage A), the downfall of lava blocks onto the slope (stage B), and the generation of pyroclastic flow (stage C). This process may be described by a function representing the single-force system that is shown in Fig. 11.21



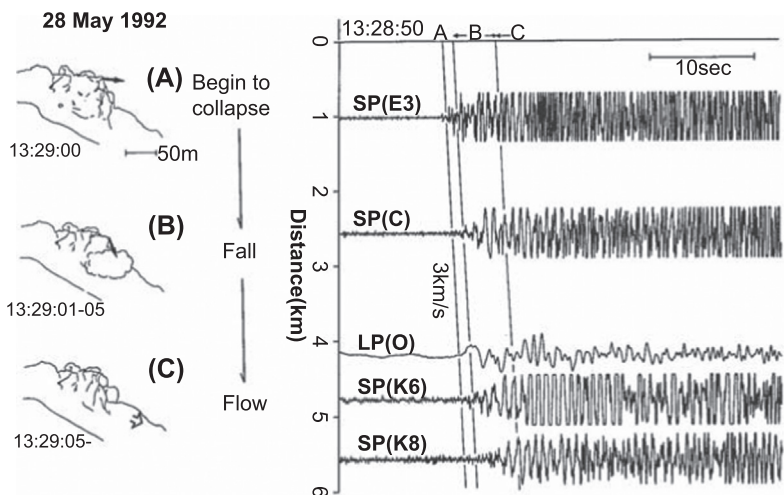


**FIGURE 11.20** Illustrations of the dome collapse sequence at Unzen volcano, Japan (left) and the corresponding vertical components of seismic signals (right). A sketch of the dome from the seismic station, situated about 3 km to south of the dome, is shown based on a video record taken by a portable 8 mm video recorder. SP, the short-period records; LP, the long-period records. The records are arranged according to the horizontal distance from the source. E3, C, O, K6, and K8 are seismic stations. From Yamasato, H., 1997. *Quantitative analysis of pyroclastic flows using infrasonic and seismic data at Unzen volcano, Japan. Journal of Physics of the Earth* 45, 397–416.

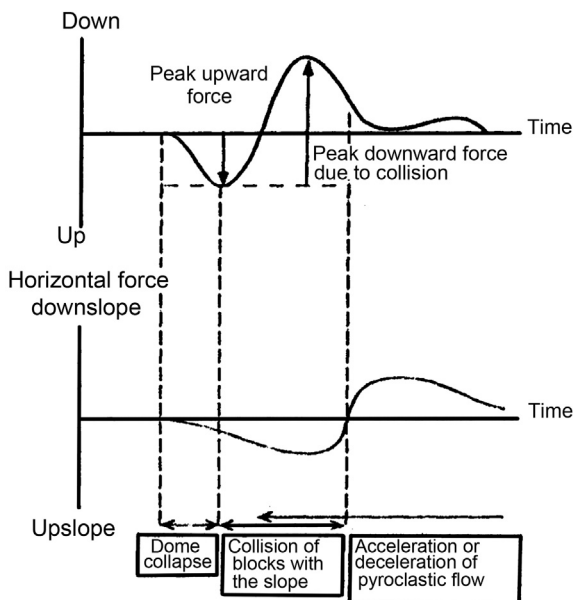


**FIGURE 11.21** Schematic source-time functions of the vertical and horizontal components of a single-force exciting pyroclastic flow according to the dome collapse sequence shown in Fig. 11.20. From Uhira, K., Yamasato, H., Takeo, M., 1994. *Source mechanism of seismic waves excited by pyroclastic flows observed at Unzen volcano, Japan. Journal of Geophysical Research* 99, 17757–17773.

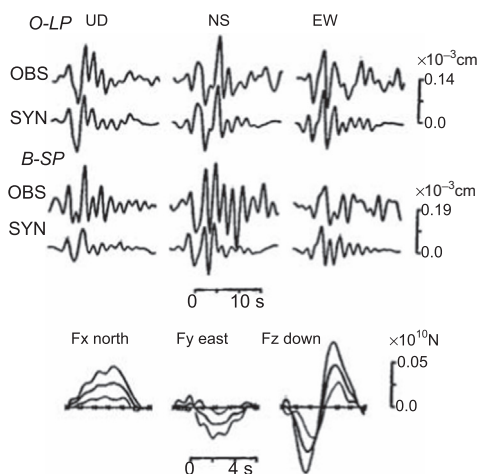


**FIGURE 11.22** Comparison of observed (OBS) and synthetic (SYN) waveforms of a seismic signal produced by the June 22, 1991 pyroclastic flow at Unzen volcano, Japan. The inversion was performed with a source-time function of the type shown in Fig. 11.21. The single-force components  $F_x$ ,  $F_y$ , and  $F_z$  used for the inversion are also shown. The inversion was realized for the short-period record of station B and the long-period record of station O. From Uhira, K., Yamasato, H., Takeo, M., 1994. Source mechanism of seismic waves excited by pyroclastic flows observed at Unzen volcano, Japan. *Journal of Geophysical Research* 99, 17757–17773.

(Uhira et al., 1994). The use of the shape of this function for the construction of a seismic source-time function allows the calculation of synthetic seismograms that reproduce the observed seismic signals associated with the propagation of the pyroclastic flows (Fig. 11.22).

An inversion gives the magnitudes of the vertical component of the single force corresponding to the collision of fragments with the slope and the horizontal component representing the deceleration of the flowing fragments along the slope.

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## Chapter 12

# Volcanic Tremor

Volcanic tremor is the very popular seismic signal recorded at volcanoes. Physical processes generating volcanic tremors involve complex interactions of magmatic fluids with the surrounding rocks. Frequently, the seismic signals of volcanic tremor precede and/or accompany most volcanic eruptions. They may be strong enough to be felt near the crater of a volcano (McNutt, 1992). Here, the general characteristics of volcanic tremor and its relationship with volcanic processes are given.

### 12.1 SEISMOGRAMS AND SPECTRA

A volcanic tremor represents a continuous record of seismic signals that may continue for minutes as an isolated tremor or for hours to months as a continuous tremor. Fig. 12.1 shows the portions of volcanic tremors recorded at Unzen volcano on November 17, 1990, just before the beginning of the eruption (this eruption is described in Section 7.1.3). An isolated tremor related to the beginning of the first eruptive event occurred at 03:22 and was followed by a continuous tremor after 03:35. The amplitude of the continuous tremor declined monotonously over a day (Shimizu et al., 1992; Nakada et al., 1999).

Volcanic tremors differ in their spectral content. The existence of two types of volcanic tremor, harmonic (or monotonic), with a fairly uniform period of seismic signals, and nonharmonic (or spasmodic), with a greater range in this respect, was found in Hawaii in 1918 by Jaggard (1920). Later the existence of these two types of volcanic tremor was recorded at many volcanoes (e.g., Sakurajima, Kawakatsu et al., 1992; Arenal, Davi et al., 2012). Seismic signals and spectral characteristics of harmonic and nonharmonic tremors recorded at Volcán de Colima, Mexico during the 2002 activity at 1.7 km from the crater (Zobin et al., 2016) are shown in Fig. 12.2. The harmonic tremor (Fig. 12.2A–C) is characterized by the linear parallel lines on the spectrogram and the one fundamental peak Fourier spectrum at 1.6 Hz. The nonharmonic tremor spectrogram (Fig. 12.2C and D) contains energy over a wide range of frequencies, and its Fourier spectrum has many individual peaks. The spectrogram represents the sequence of the amplitude bursts at different frequency levels.

Volcanic tremor may be represented by surface waves or body waves (Figs. 12.3 and 12.4).

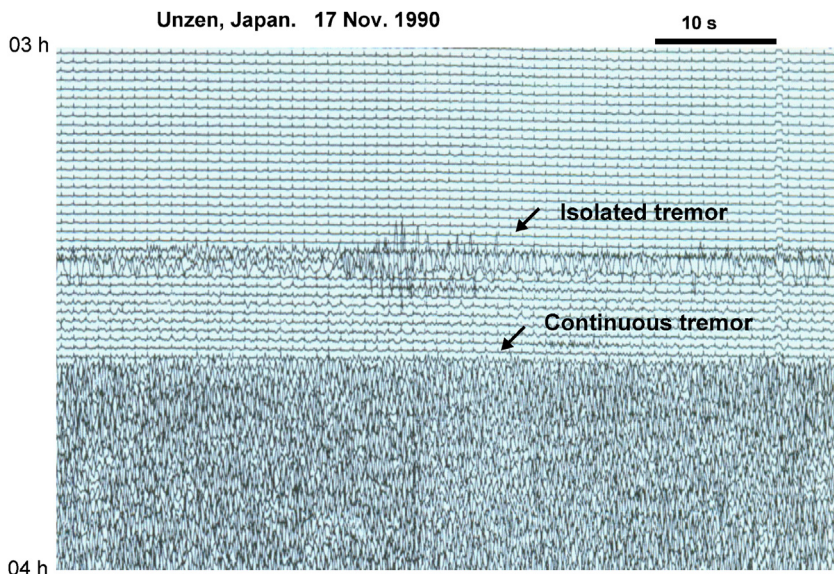


**FIGURE 12.1** Isolated and continuous volcanic tremor recorded at Unzen volcano, Japan on November 17, 1990, with a short-period instrument at a distance of 3.3 km from the crater of the volcano. The beginning of isolated tremor record is supposed to mark the beginning of the eruption of volcano (see Section 7.1.3). *Modified from Shimizu, H., Umakoshi, K., Matzuwo, N., Ohta, K., 1992. Seismological observations of Unzen volcano before and during the 1990–1992 eruption. In: Tanagi, Y., Okada, H., Ohta, K. (Eds.), Unzen Volcano: The 1990–1992 Eruption. The Nishinippon and Kyushu University Press, Fukuoka, pp. 38–43.*

## 12.2 LOCATION OF VOLCANIC TREMOR

The sources of volcanic tremor are located at shallow depths beneath the crater floor.

### 12.2.1 Oshima Volcano, Izu Islands

[Yamaoka et al. \(1991\)](#) estimated the position of the source of the wide-frequency tremor that preceded a small explosive eruption on November 16, 1987, of Oshima volcano from the cross-correlate analysis of the 0.7 Hz component and the propagation time of the tremor recorded by a dense 13-station seismic network ([Fig. 12.5](#)). They obtained a source location at a depth of about 350 m above sea level (a.s.l.), which is slightly deeper than the crater bottom before the 1986 Oshima eruption (see Section 5.2.1).

### 12.2.2 Etna Volcano, Sicily

The tremor source during the 2008–09 Etna eruption was estimated by two methods: using the seismic amplitude decay and by the comparison of the



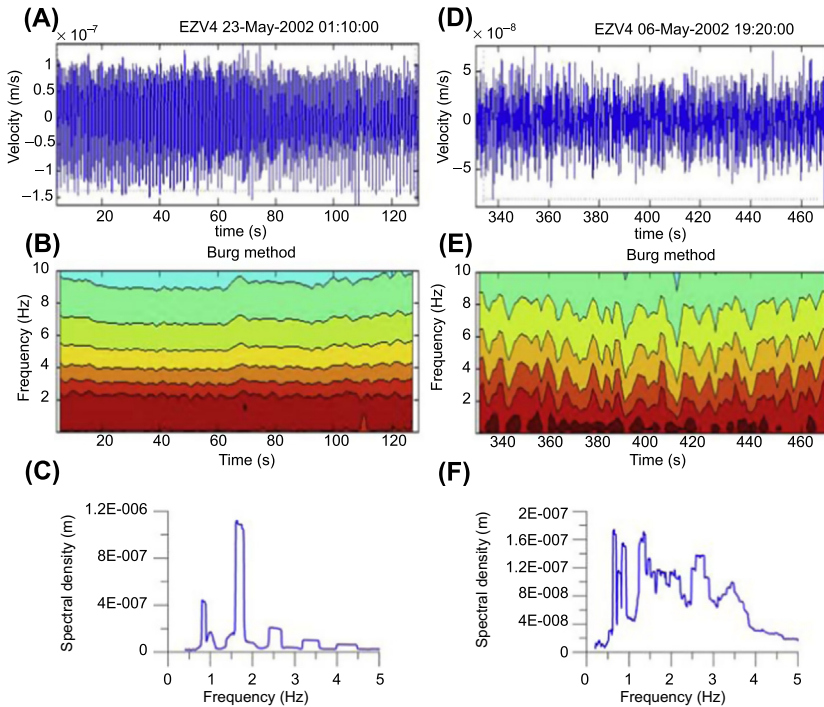


**FIGURE 12.2** Examples of harmonic (May 23, 01:10) and nonharmonic (May 6, 19:20) volcanic tremor recorded at Volcán de Colima, Mexico and presenting in the short-period vertical component signals (A, D) recorded at a distance of 1.7 km from the crater, their spectrograms (B, E) and Fourier spectra (C, F), respectively. In spectrograms, the darker colors indicate larger amplitudes of spectral density. Spectrograms were calculated with a window length of 256 counts and an overlap of 128 counts. *Courtesy Colima Volcano Observatory.*

seismic signal envelopes (Cannata et al., 2010). Signals were recorded by 16 stations, belonging to the broadband permanent seismic network and located at distances ranging between 1.5 and 9 km from the center of the summit area (maximum elevation of 3300 m a.s.l.). The first method assumed the propagation in a homogeneous medium and seismic amplitude decay with distance. Three different frequency bands, namely 1–5, 5–10, and 10–15 Hz, were analyzed (Fig. 12.6). In the first case (Fig. 12.6A), the tremor sources were located in a volume between the eruptive fissure and the South–East crater (SEC) at an altitude of 1.0–2.8 km a.s.l. The source of the seismic signal, filtered in the bands 5–10 and especially 10–15 Hz (Fig. 12.6B and C), was located below the SEC at a very shallow depth (about 3 km a.s.l.). The second method, based on the comparison of the seismic signal envelopes, gave the locations which are generally consistent with the locations obtained by the first method (Fig. 12.6A–C). At the same time, a regular shift is seen in these source estimations compared with the sources obtained by the first method.

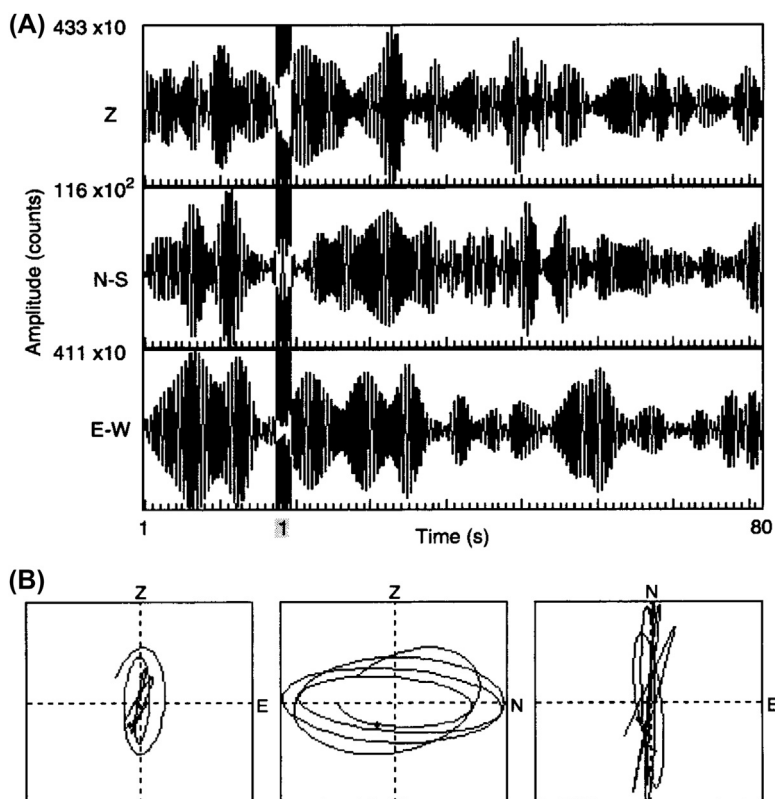


**FIGURE 12.3** Example of particle-motion analysis on a time series of tremors recorded during 1995 at Stromboli volcano, Italy at a distance of 1.8 km from the crater. (A) Three-component short-period seismic records filtered with a Butterworth bandpass between 1.55 and 2 Hz; (B) particle motion. The north–south-directed particle motion with subhorizontal incidence indicates a source close to the *S*- and the *P*-wave nature of volcanic tremor. From Falsaperla, S., Langer, H., Spampinato, S., 1998. Statistical analysis and characteristics of volcanic tremor on Stromboli volcano (Italy). *Bulletin of Volcanology* 60, 75–88.

### 12.2.3 Arenal Volcano, Costa Rica

Davi et al. (2012) performed the inversion of the 15 tremor events recorded in February 2005 on the andesitic stratovolcano Arenal by a temporal broadband seismic network. A simultaneous grid search for the source mechanisms and location was performed for each event. A common source located 100 m below the crater summit (Fig. 12.7) was obtained to evaluate the joint probability density function of the misfit values for all the events. For this common source point, solutions for moment tensor (MT) components and moment tensor components plus single forces (MT + SF) were evaluated. Fig. 12.8 shows the example of a moment tensor solution in



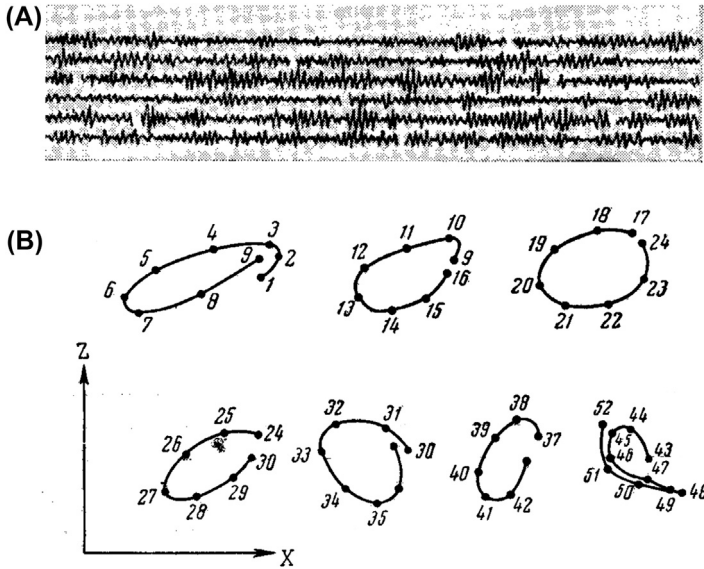


**FIGURE 12.4** Example of particle-motion analysis on a time series of tremors recorded during 1961 at Klyuchevskoy volcano, Kamchatka at a distance of 9 km from the crater. (A) the E–W component of short-period seismic records; (B) particle motion. The ellipse-shaped particle motion indicates the Raleigh-wave nature of volcanic tremor. From Tokarev, P.I., 1966. *Eruptions and Seismic Regime of Volcanoes of Klyuchevskaya Group*. Nauka, Moscow (in Russian). 118 p.

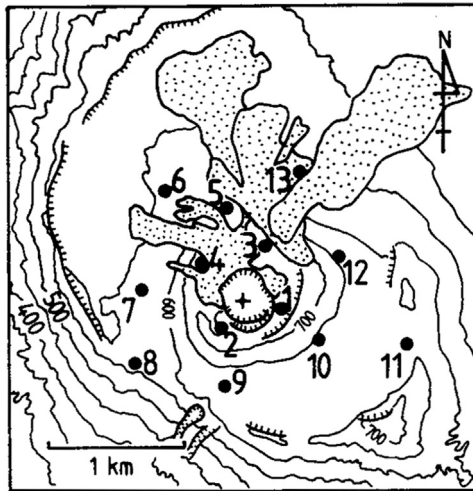


**FIGURE 12.5** Source location (+) of volcanic tremor in the summit area of Oshima volcano. Seismic network is shown by the black circles with numbers. From Yamaoka, K., Oikawa, J., and Ida, Y., 1991. An isotropic source of volcanic tremor—observation with a dense seismic network at Izu-Oshima volcano, Japan. *Journal of Volcanology and Geothermal Research* 47, 329–336.

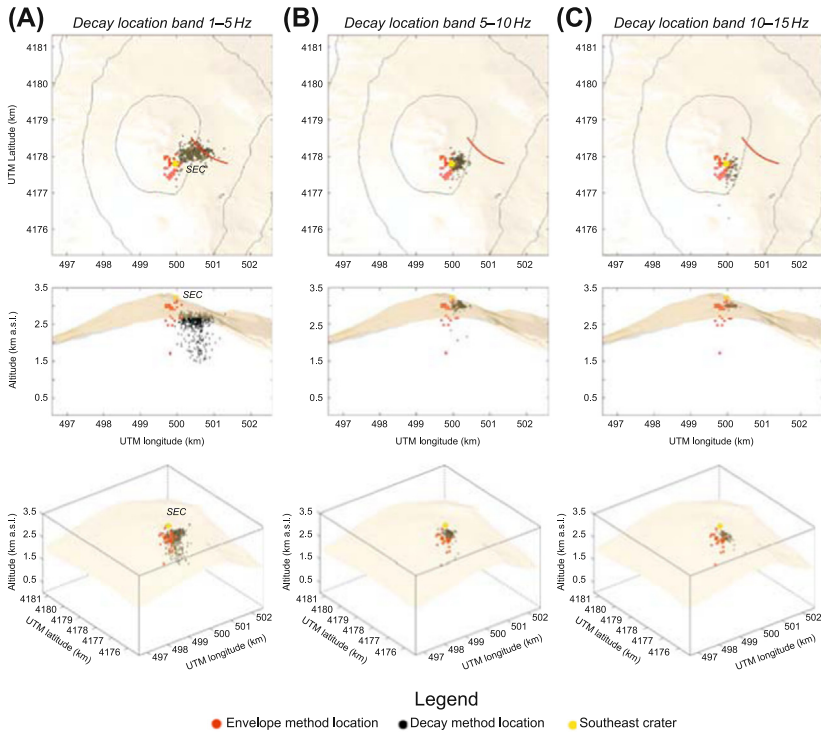


**FIGURE 12.6** Source locations of Etna volcanic tremor obtained by amplitude decay in the frequency bands 1–5 (A), 5–10 (B), and 10–15 (C) Hz on August 29 and October 2 to October 3, and by the comparison of the seismic signal envelopes during August 27 to September 3 and October 2, 2008, to October 3, 2008. *EF*, Eruption fissure; *SEC*, southeast crater. From Cannata, A., Di Gracia, G., Montalvo, P., Ferrari, P., Nunnari, G., Patanè, D., et al., 2010. *New insights into banded tremor from the 2008–2009 Mount Etna eruption. Journal of Geophysical Research* 115, B12318.

which the obtained MT + SF and MT solutions are plotted for the event recorded on February 13, 2005. These results are similar for the 15 events. The decomposition method indicates solutions with  $50\% \pm 5\%$  of isotropic component,  $35\% \pm 12\%$  of compensated linear vector dipole (CLVD), and  $15\% \pm 17\%$  of DC for all the events. The source geometry for the tremor signals is represented by a subhorizontal crack dipping  $\sim 20^\circ$  toward the north–northeast.

### 12.3 RELATIONSHIP BETWEEN THE INTENSITY OF VOLCANIC TREMOR AND VOLCANIC EVENTS

For proper comparison between the tremor intensity and eruption characteristics at different volcanoes, the standardization of tremor amplitudes to a

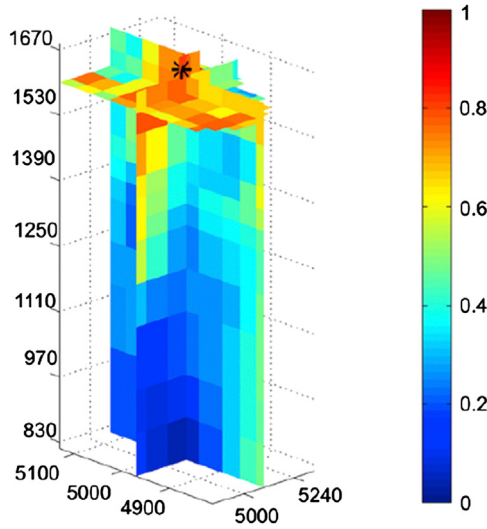


**FIGURE 12.7** Normalized joint probability density function of the source position for the 15 tremor events at Arenal volcano for moment tensor (MT) components without single forces. The dark star indicates the chosen source position. The distances are expressed in meters. From Davi, R., O'Brien, G.S., De Barros, L., Lokmer, I., Bean, C.J., Lesage, P., et al., 2012. Seismic source mechanisms of tremor recorded on Arenal volcano, Costa Rica, retrieved by waveform inversion. *Journal of Volcanology and Geothermal Research* 213–214, 1–13.

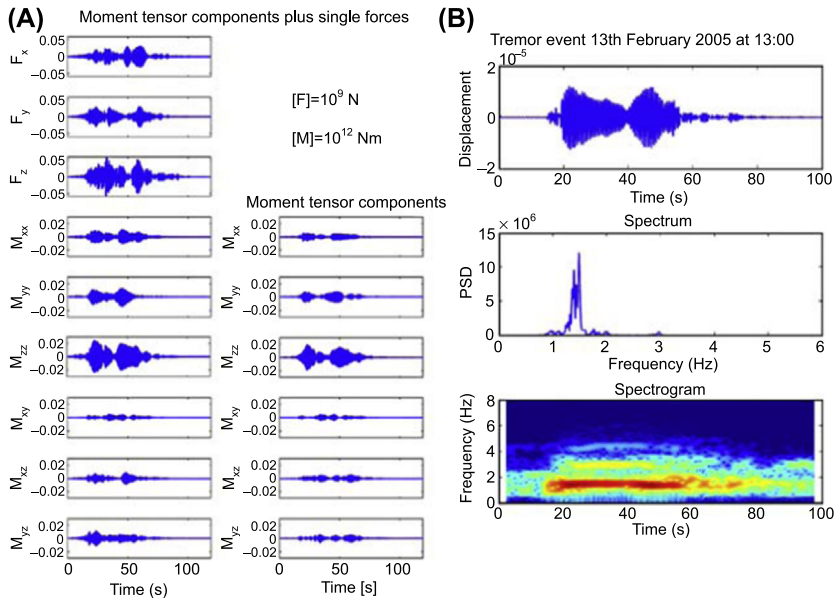


**FIGURE 12.8** Moment tensor solution of a tremor event recorded on February 13, 2005. (A) Moment tensor inversion solution for MT + SF (left) and MT (right). (B) From the top to bottom of the waveform are the spectrum and the spectrogram, respectively. MT, Moment tensor components; MT + SF, moment tensor components plus single forces. From Davi, R., O'Brien, G.S., De Barros, L., Lokmer, I., Bean, C.J., Lesage, P., et al., 2012. Seismic source mechanisms of tremor recorded on Arenal volcano, Costa Rica, retrieved by waveform inversion. *Journal of Volcanology and Geothermal Research* 213–214, 1–13.

common reference is used. McNutt (1992) proposed the following formula for estimation of the reduced displacement (RD) of tremor signals:

$$\text{RD (reduced displacement)} = \frac{(A\sqrt{\lambda r})}{(2\sqrt{2M})} \quad (12.1)$$

Here  $A$  is a peak-to-peak amplitude in cm,  $r$  is the distance from the source to the seismic station in cm,  $M$  is seismograph magnifications at the tremor frequency,  $\lambda$  is wavelength in cm, and  $2\sqrt{2}$  is the root-mean-square amplitude correction.

Fig. 12.9 shows a comparison between volcanic tremor amplitude and video camera frames of the May 29, 2000 lava fountain episode at Etna volcano (Alparone et al., 2003). A good correlation between the amplitudes of volcanic tremors and the height of lava fountains and pyroclastic material was noted for many volcanoes (Figs 12.10 and 12.11). The data obtained for two volcanoes (Kilauea, Hawaii, and Pavlof, Alaska) overlap (Fig. 12.10), suggesting similar relationships for both volcanoes. Tokarev et al. (1984) showed that the time of output of incandescent material coincided with the time of origin of an intensive seismic impulse in the sequence of volcanic tremors recorded during the November 1976 Strombolian explosions at the southern side of New Tolbachik volcanoes in Kamchatka (Fig. 12.11A).

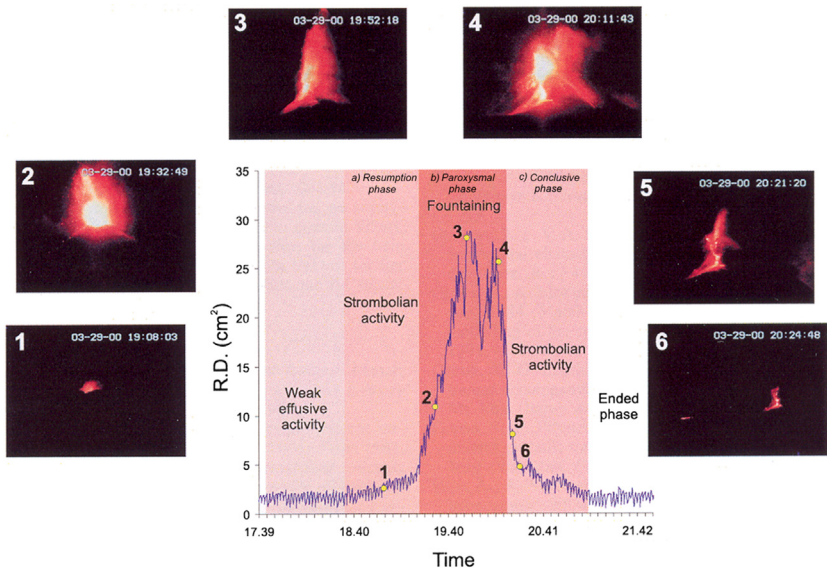


**FIGURE 12.9** Comparison between volcanic tremor amplitude and video camera frames of the March 29, 2000 lava fountain episode at Etna volcano. From Alparone, S., Andronico, D., Lodato, L., Sgroi, T., 2003. Relationship between tremor and volcanic activity during Southeast Crater eruption on Mount Etna in early 2000. *Journal of Geophysical Research* 108, 2241.

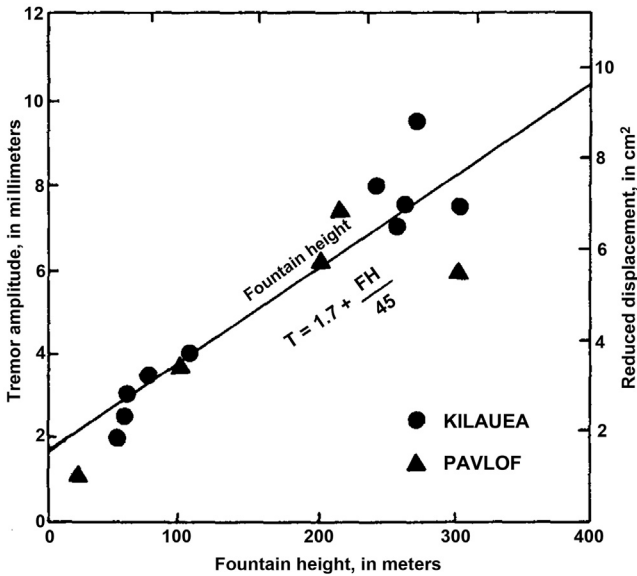


**FIGURE 12.10** Relationship between the fountain height at basaltic volcanoes Kilauea, Hawaii (round dots) and Pavlov volcano, Alaska (triangles) and tremor amplitude. Both data sets were normalized to units of reduced displacement. The line represents a lineal regression fit to Kilauea data. From McNutt, S.R., Miller, T.P., Taber, J.J., 1991. Geological and seismological evidence of increased explosivity during the 1986 eruptions of Pavlov volcano, Alaska. *Bulletin of Volcanology* 53, 86–98.

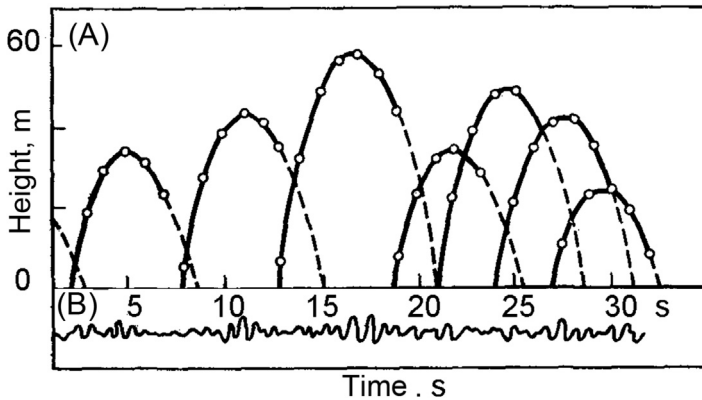


**FIGURE 12.11** Dependence of the amplitude of volcanic tremor upon the height of pyroclastic material rising out during the explosions at New Tolbachik volcanoes, Kamchatka. The shape of pyroclastic cones (A) and a seismogram (vertical component) of volcanic tremor (B) were recorded simultaneously by filming and by short-period seismometer, respectively at a distance of about 3 km from the site of the eruption. The seismogram is moved to the left for 2.5 s to correct for the seismic wave propagation from the site of eruption to seismic station. From Tokarev, P.I., Gordeev, E.I., Farberov, A.I., 1984. Volcanic tremor. In: Fedotov, S.A., (Ed.), *Large Tolbachik Fissure Eruption*. Nauka, Moscow, pp. 448–471 (in Russian).

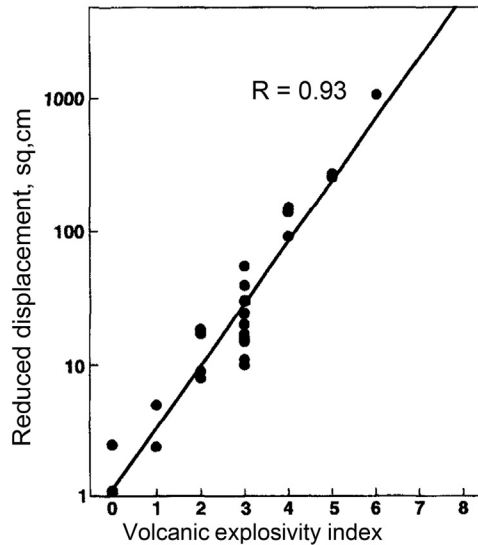


**FIGURE 12.12** Volcanic tremor reduced displacement versus volcano explosivity index (VEI; Newhall and Self, 1982).  $R$  is the correlation coefficient. The least square regression line is shown. From McNutt, S.R., 1994. Volcanic tremor amplitude correlated with eruption explosivity and its potential use in determining ash hazards to aviation. In: Volcanic Ash and Aviation Safety. U.S. Geol. Surv. Bull. 2047, 377–385.

The amplitude of volcanic tremor was well correlated with the height of pyroclastic material rising out of the volcanic crater. With an increase in the explosion frequency, the tremor impulses interfere with the decrease of the amplitude (Fig. 12.11B).

A more general relationship between the maximum amplitude of volcanic tremor and the volcanic explosivity index (VEI; see Section 1.1.1) was obtained by McNutt (1994) based on the data from 21 eruptions at 14 volcanoes (Fig. 12.12). A lineal regression of amplitudes (in RD units) plotted versus VEI yielded the following regression:

$$\log RD(\text{cm}^2) = 0.48(\text{VEI}) + 0.08(12.2) \quad (12.2)$$

## 12.4 VOLCANIC TREMOR IN ERUPTIVE PROCESS

Volcanic tremor reflects many important variations in eruptive activity. Some cases of these relationships between the amplitudes and spectral content and the eruption dynamics are explained in the following sections.

### 12.4.1 Etna Volcano, Sicily

Gresta et al. (1991) demonstrated a good relationship between the course of the 1984 activity at Etna volcano, Sicily, and the variations in the amplitude

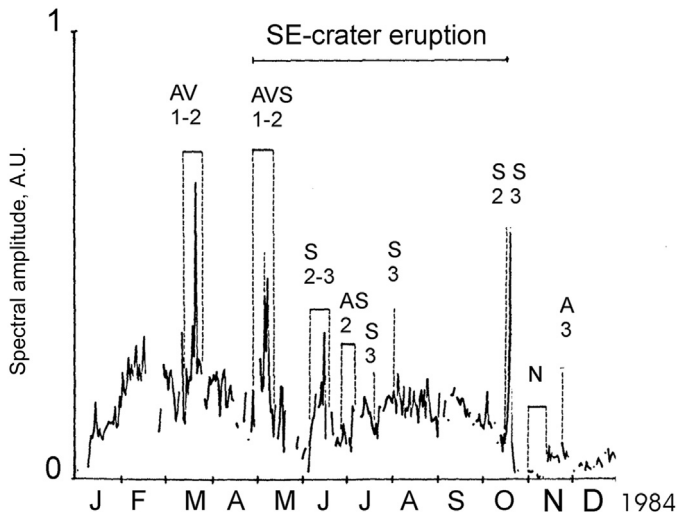


**FIGURE 12.13** Amplitude variations of spectral peak at 1.2 Hz of volcanic tremors recorded at Etna volcano during its 1984 activity. The types of volcanic activity are marked by A (ash ejection), V (emission of vapor and/or gas), and S (Strombolian activity). The intensity of eruptive activity is shown by 1 (moderate), 2 (strong), and 3 (very strong). N corresponds to the absence of any activity. AU, the arbitrary units. After Gresta, S., Montalto, A., Patané, G., 1991. Volcanic tremor at Mount Etna (January 1984–March 1985): its relationship to the eruptive activity and modeling of the summit feeding system. *Bulletin of Volcanology* 53, 309–320.

of volcanic tremors. They showed that this relationship is clearer for the spectral amplitudes than for overall amplitudes. The most sensitive were the amplitudes of a 1.2 Hz peak (Fig. 12.13); it was the only one that displayed coincidence with the remarkable events of volcanic activity at all four summit craters of the volcano. At the same time, it can be seen that an increase in the spectral amplitudes of the tremor does not depend on the type or intensity of volcanic activity; it may only indicate a surface eruptive manifestation. The spectral amplitudes of the peaks of 0.95, 1.45, 1.65, 1.8, and 2.4 Hz, which were also dominant during the eruptive activity, were characterized by a lower sensibility to eruptive manifestations.

#### 12.4.2 Pavlof Volcano, Alaska

The eruption was continuous from September 16, 1996, to December 4, 1996, declined, and then had two stronger bursts on December 10 to December 12 and December 27 to December 29 (McNutt, 2000). Fig. 12.14 demonstrates the ability of tremors to reflect the explosive phases of the eruption by an increase in amplitude. The two largest peaks in amplitude correspond to the stronger December bursts.



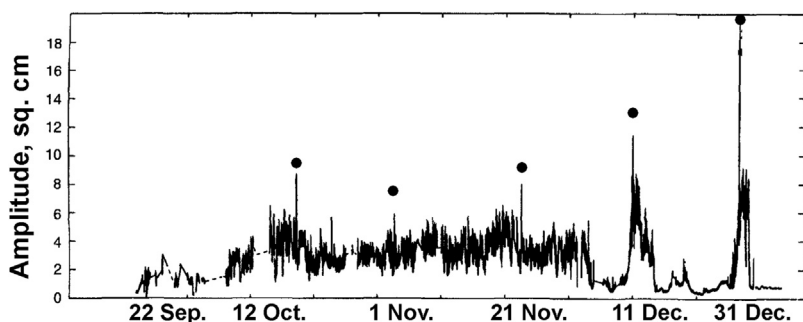


**FIGURE 12.14** Variation of the normalized amplitude of volcanic tremor during the 3.5 months of the 1996 Pavlof eruptions. Stronger tremors accompanied more explosive phases (shown by filled points) on October 18, November 4 and November 23, and December 10 and December 27. From McNutt, S.R., 2000b. *Seismic monitoring*. In: H. Sigurdsson, (Ed.), *Encyclopedia of Volcanoes*. Academic Press, San Diego, pp. 1095–1120.

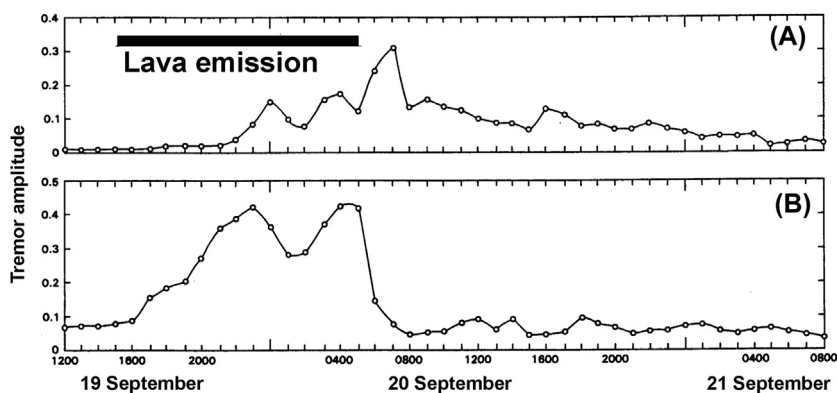


**FIGURE 12.15** Variation in the tremor peak-to-peak amplitudes (in micrometers) during and after the September 1984 East-Rift lava eruption at Kilauea volcano, Hawaii at the volcanic summit (A) and at a distance of 6 km west from the site of lava emission in the East-Rift zone (B). The period of lava emission is shown. From Koyanagi, R., Chouet, B., Aki, K., 1987. *Origin of volcanic tremor in Hawaii. Part I*. In: Decker, R.W., Wright, T.L., Stauffer, P.H. (Eds.), *Volcanism in Hawaii*. U.S. Geol. Surv. Prof. Paper 1350, 1221–1257.

### 12.4.3 Kilauea Volcano, Hawaii

The 1984 East-Rift eruption, which produced about  $11 \times 10^6 \text{ m}^3$  of lava, began on September 19 at 16 hours and continued for 13.5 hours. This eruption was accompanied by harmonic volcanic tremor recorded at a distance of 6 km west from the site of the eruption and at the summit of the Kilauea volcano. Fig. 12.15 shows the variations of the tremor amplitude recorded at the volcanic summit and in the East-Rift zone, near the site of the eruption. The tremor near the site of the eruption was more intensive than at the summit of Kilauea. The tremor began with the start of the eruption, gradually



reached its maximum value over 7 hours, and stayed at this level up to the end of the eruption. A rapid decrease in tremors accompanied the end of sustained high fountaining. The summit tremor began 6 hours after the start of the eruption and reached its maximum amplitude 2 hours after the cessation of the lava emission (Koyanagi et al., 1987).

#### 12.4.4 Klyuchevskoy Volcano, Kamchatka

The September 8, 1994 summit eruption of Klyuchevskoy volcano was preceded and accompanied by intense volcanic tremor (Ivanov, 2011). The eruption began with lava fountaining and ash plumes and continued for about 1 month. It culminated in the paroxysmal explosion on October 1 with the ash column estimated at 15–20 km altitude a.s.l. and extended > 100 km SE.

Fig. 12.16A shows the variations in the intensity of seismic signals, expressed in the relation of the mean daily amplitude of tremor to the mean period of tremor signals during this day and recorded at a distance of 11 km from the crater (Ivanov, 2011). It is seen that the intensity of the tremor sharply increased with the beginning of the eruption on September 8. The gradual increase in tremor intensity continued up to September 12. On September 12, a large gas and ash explosion occurred. The ash plume reached an altitude of about 8 km a.s.l. The explosion was accompanied by a 1-km-long lava flow on the SW slope of the volcano; mudflows were also noted. Beginning from September 12, the intensity level of continuous volcanic tremor was high and relatively stable till September 20. During this period, gas and ash bursts rose 500–700 m above the crater. The eruption column reached 1.5–2.0 km above the crater. Lava flows extruding from two vents 200 m below the crater rim had moved down to 2800 m elevation on the NW and SW flanks. Phreatic explosions were occurring at the contact of the NW lava flow and the glacier. Lava fountains in the central crater reached heights of 300–500 m.

From September 20, a slow growth of tremor intensity began (Fig. 12.16A). During this period, gas and ash bursts increased in height to 800–1000 m above the crater. The eruption columns continued to reach ~2 km above the crater. Lava flows on the NW and SW flanks remained active, and fountains in the central crater increased to a height of 500–700 m. Two days before the paroxysmal explosion of October 1, the level of volcanic tremor began to increase significantly up to the moment of this explosion. Volcanic tremor was continuous with a maximum amplitude of 8.4  $\mu\text{m}$ . After the paroxysmal explosion, volcanic tremors practically disappeared.

The detailed analysis of the tremor intensity shows that about one week before the September 8 eruption there were recorded the precursors to this eruption (Fig. 12.16B). The low-level background tremor, which began about 2 months before the eruption (stage I in Fig. 12.16B), gradually increased from September 2 to September 8 (stage II). Then it sharply increased during 8–12 hours before the eruption (stage III; Ivanov, 2011).

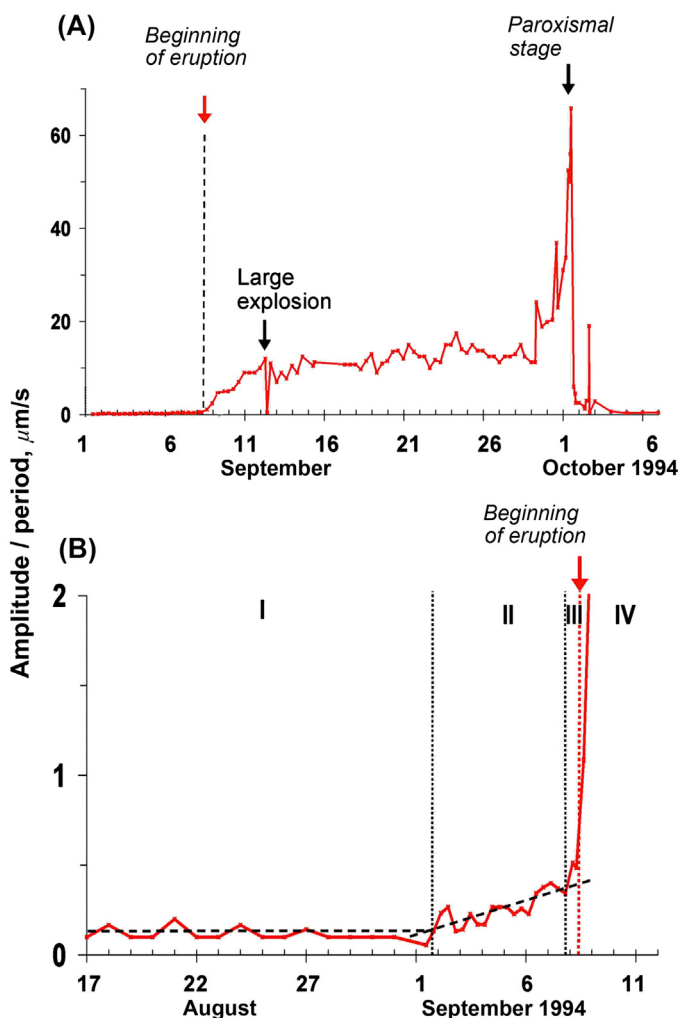


**FIGURE 12.16** Variations in the intensity of volcanic tremor before and during the 1994 eruption of Klyuchevskoy volcano, Kamchatka, from September 1994 to October 1994 (A) and the detailed representation of preliminary variations of tremor intensity from August 17, 1994, to September 12, 1994 (B). In (B), I is the stage of the background level of tremor; II, the stage of linear increase in tremor intensity; III, the stage of exponential growth of tremor intensity coinciding with increase of degassing activity of the volcano; and IV, eruption stage. After Ivanov, V.V., 2011. *Volcanic tremor as a precursor to the summit eruptions of Klyuchevskoy volcano, Kamchatka and a possible mechanism of volcanic tremor. Proc. Conf. Volcanism and Related Processes. Institute of Volcanology and Seismology, Petropavlovsk–Kamchatsky*, pp. 42–46 (in Russian).

### 12.4.5 Shishaldin, Unimak Island

Shishaldin, a basaltic stratovolcano, is the most active volcano on Unimak Island, Aleutian Arc. Its 1999 eruption that began in February culminated with sub-Plinian explosions on April 19 and April 23. Volcanic tremor was observed at all stages of the eruption (Thompson et al., 2002). Fig. 12.17A shows the temporal variations in the amplitudes of RD of the tremor signals during the period from April 4, 1999, to April 24, 1999. A tremor that was

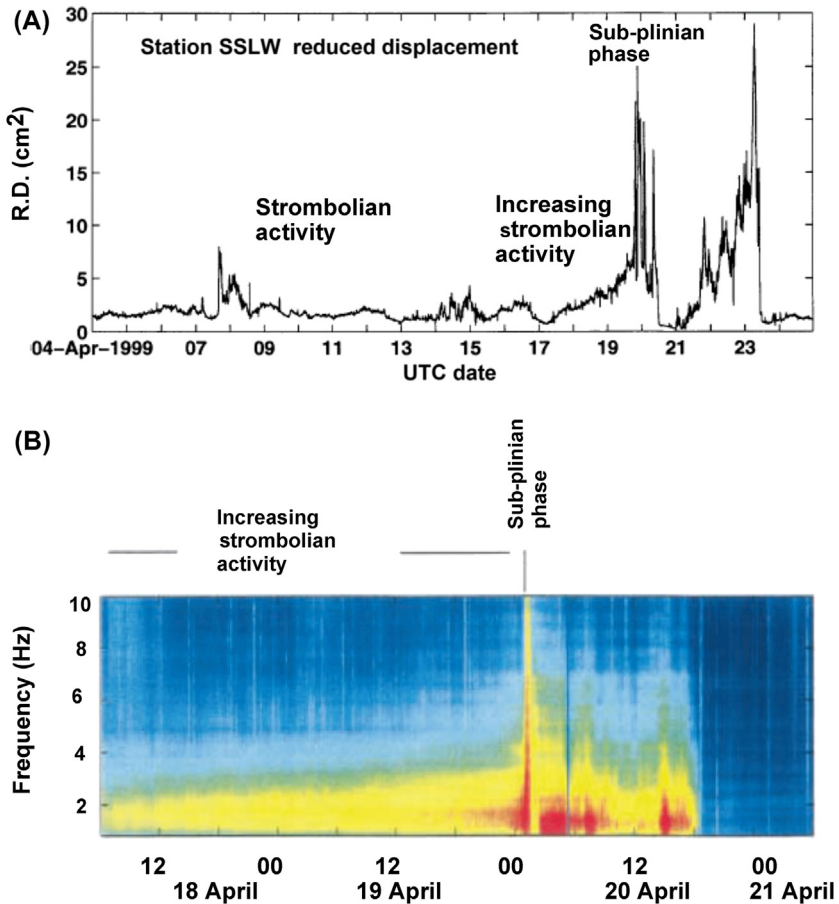


**FIGURE 12.17** (A) Reduced displacement (RD) of volcanic tremor recorded from April 4, 1999, to April 24, 1999, at Shishaldin Volcano. (B) Spectrogram of the volcanic tremor recorded from April 18, 1999, to April 21, 1999, before and after the sub-Plinian phase of eruption. Data are from short-period station SSLW situated at a distance of 10 km from the crater. From Thompson, G., McNutt, S.R., Tytgat, G., 2002. Three distinct regimes of volcanic tremor associated with the eruption of Shishaldin Volcano, Alaska 1999. *Bulletin of Volcanology* 64, 535–547.

more or less stationary at the beginning of April (RDs between 2 and 4 cm<sup>2</sup>), with the occurrence of small Strombolian eruptions from April 7 to April 8 had increased abruptly with RDs of about 8 cm<sup>2</sup>. The sub-Plinian VEI = 3 eruptions occurred on April 19, lasting 50–80 minutes, and were followed by three additional strong Strombolian phases. RDs began to grow gradually during increasing Strombolian activity beginning from April 17 and reached 23 cm<sup>2</sup> on April 19 and 43 cm<sup>2</sup> on April 23.

The time series of amplitude spectra of tremor just before the culminated sub-Plinian explosion of April 19 (Fig. 12.17B) shows the evolution in the tremor spectral characteristics during the transition from background volcanic unrest to the eruption of high intensity. The spectrogram demonstrates the temporal variations in the frequency content of tremors during the period of eruption between April 18 and April 21. Tremor frequencies were dominantly 1–2.5 Hz during the period of increasing Strombolian activity. Several hours before the sub-Plinian phase on April 19 the tremor with a peak frequency of 1.3 Hz intensified. During the sub-Plinian phase, the tremor became more broadband in spectral content with the peak frequency increased to 4–8 Hz. After the April 19 event, alternating bands of high- and low-amplitude tremors occurred.

## 12.5 SPECIAL CASES OF VOLCANIC TREMOR

### 12.5.1 Isolated Tremor

Isolated tremor is a spasmodic seismic event and its envelope is spindle or irregular shaped (see Fig. 12.1).

#### 12.5.1.1 *Teishi-Knoll Volcano*

Large-amplitude isolated volcanic tremor was observed before the 1989 Teishi Knoll, Izu islands eruption. Fig. 12.18 shows the position of these signals in the total sequence of seismotectonic events preceding the eruption (Ukawa, 1993). The sequence commences with a swarm of volcano-tectonic earthquakes (see Section 5.5.1) that continued for about 7 days from July 4 and was accompanied by a large increase in crustal tilt. After the cessation of the volcano-tectonic activity and during the maximum tilt, a sequence of low-frequency earthquakes occurred. At the coda of the sequence of low-frequency earthquakes, the continuous tremor began. It culminated with the large-amplitude isolated tremor that took place for about 1 hour and its conclusion was coincident with the beginning of the submarine eruption.

Figs 12.19 and 12.20 show the waveforms of the volcanic tremor and their spectra. In the figures, F1 and F2 indicate the episodes of continuous tremor occurring on July 11 and July 12, the large-amplitude tremor of July 13 is shown as the main phase, and the isolated events are shown in the lower part. The power spectra of these events demonstrate that every tremor has its spectral peak at around 0.8 Hz. The spectral shape of the main phase

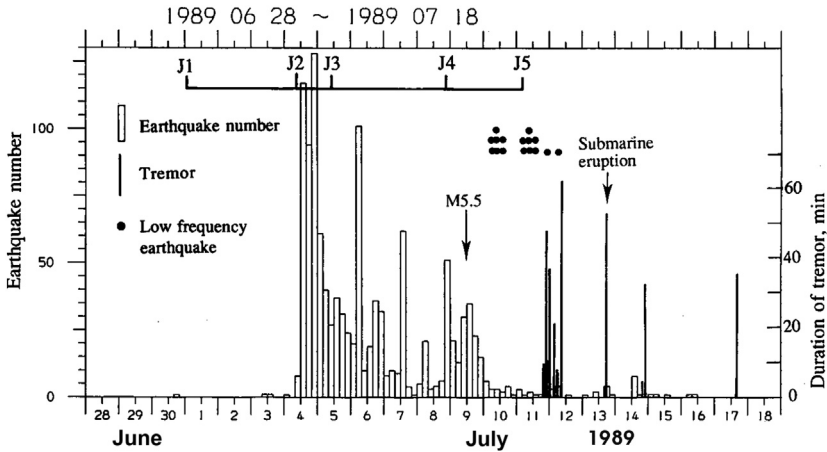


**FIGURE 12.18** Sequence of the seismotectonic events associated with the 1989 Teishi Knoll, Izu islands eruption. Earthquake number is shown for every 4 h; the epochs J1–J5 reflect the increase of crustal tilt from J2 to J5 that exceeded 20 mrad. *From Ukawa, M., 1993. Excitation mechanism of large-amplitude volcanic tremor associated with the 1989 Ito-oki submarine eruption, central Japan. Journal of Volcanology and Geothermal Research 55, 33–50.*

is similar to that of the isolated events. The spectra of the continuous tremors F1 and F2 are different in shape from those of the large-amplitude tremors and are characterized by narrower peak bands. These spectral properties indicate that the tremors F1 and F2 were produced by smaller-scale events compared with the large-amplitude tremors (Hashimoto et al., 1991).

Oshima et al. (1991) considered that the continuous tremors F1 and F2 of July 11 were associated with the formation of a new volcanic knoll (cryptodome). The large-amplitude tremor of July 13 was believed to be caused by the final moments of preeruptive and eruptive magma movements. The hypocenters of the isolated events are located near the position of the submarine eruption (Goto et al., 1991).

### 12.5.1.2 New Tolbachik Volcanoes

The newborn New Tolbachik cones in Kamchatka were formed along a freshly opened fault on July 6, 1975 (see Section 5.3.1). The 16-minute record of low-frequency volcanic tremor was observed just before the beginning of the preceding swarm of volcano-tectonic earthquakes (Fig. 12.21). It represented the continuous vibrations with a gradual increase in the amplitude. Maximum amplitudes were four times larger than the initial amplitudes. After reaching a maximum, the amplitudes began to gradually decrease. With the onset of the volcano-tectonic earthquake swarm, the tremors ceased and were not observed for the following 2-week seismic activity up to the beginning of the explosive activity of the newly formed cone (Tokarev et al., 1984).

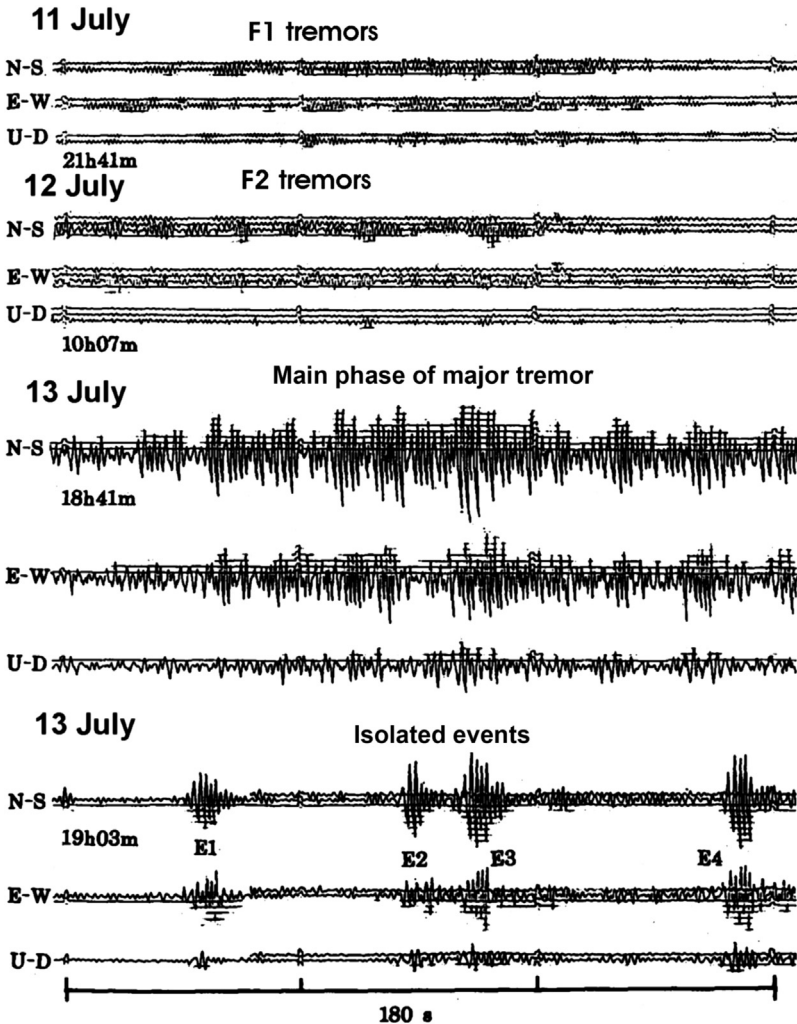


**FIGURE 12.19** Seismograms of the isolated tremors preceding the 1989 Teishi Knoll, Izu islands eruption. Recorded at a distance of about 8 km from the site of a submarine eruption. From Hashimoto, T., Maeda, K., Odaka, T., Seino, M., 1991. Waveform analysis of volcanic tremors associated with the July 1989 volcanic activity off the east coast of Izu peninsula. *Journal of Physics of the Earth* 39, 65–78.

### 12.5.1.3 Banded Tremor

A particular kind of volcanic tremor characterized by regular cyclic increases in amplitude was observed in several volcanic and hydrothermal areas (Fig. 12.22). Because of its peculiar seismic signature, forming evident stripes on seismograms, this kind of seismic signal is called a “banded” tremor (Cannata et al., 2010).

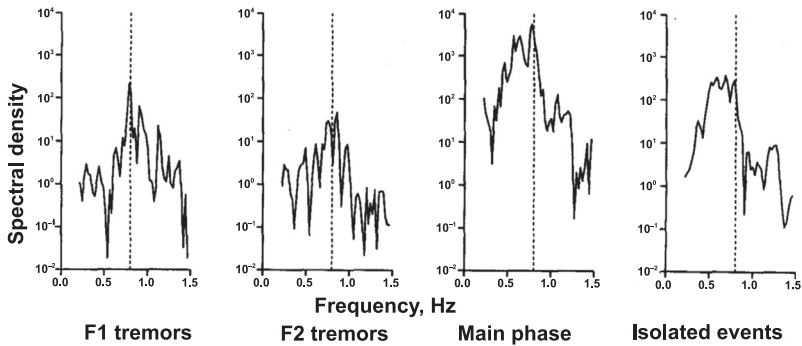


**FIGURE 12.20** Power spectral density (in  $\mu\text{m}^2/\text{s}$ ) curves of NS component of isolated tremors preceding the 1989 Teishi Knoll, Izu islands eruption shown in Fig. 12.16. Dashed lines indicate the frequency of 0.8 Hz. From Hashimoto, T., Maeda, K., Odaka, T., Seino, M., 1991. Waveform analysis of volcanic tremors associated with the July 1989 volcanic activity off the east coast of Izu peninsula. *Journal of Physics of the Earth* 39, 65–78.

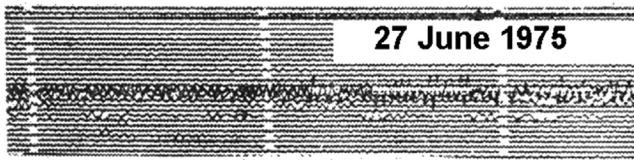


**FIGURE 12.21** Low-frequency isolated volcanic tremor recorded by a short-period seismic station at a distance of about 50 km from the site of a newborn cone of New Tolbachik volcanoes, Kamchatka, 10 days before the beginning of the July 6 eruption. Seismic record is courtesy of the Kamchatkan Methodical Seismological Department of the RAS Geophysical Survey, Russia.

#### 12.5.1.4 Miyakejima Volcano

The study of banded tremor was realized at Miyakejima volcano after its 2000 eruption (see Section 5.5.2). As was noted by Fujita (2008), the banded tremor was not observed throughout the entire period of the 2000 volcanic activity but only intermittently between 2001 and 2006. The active periods of banded tremor were occasional and each period lasted about 1–2 days on average, ranging from a few hours to 10 days. Remarkable banded tremor was recorded during the periods of April and August 2002; March, April, August, and September 2003; March–April, May–June, July–August, and November–December 2004; and April and August 2005. It seems that the banded tremor had seasonal activity. The durations of the tremor episodes in each active period were similar. The duration of signals during the spring season (March and April) was shorter than those in summer (May–August). Fig. 12.23 shows the histograms of durations and intervals observed during the whole 2001–06 period of observations and this clearly shows that there were typical values for both characteristic times about 6.5 and 25 minutes, respectively.



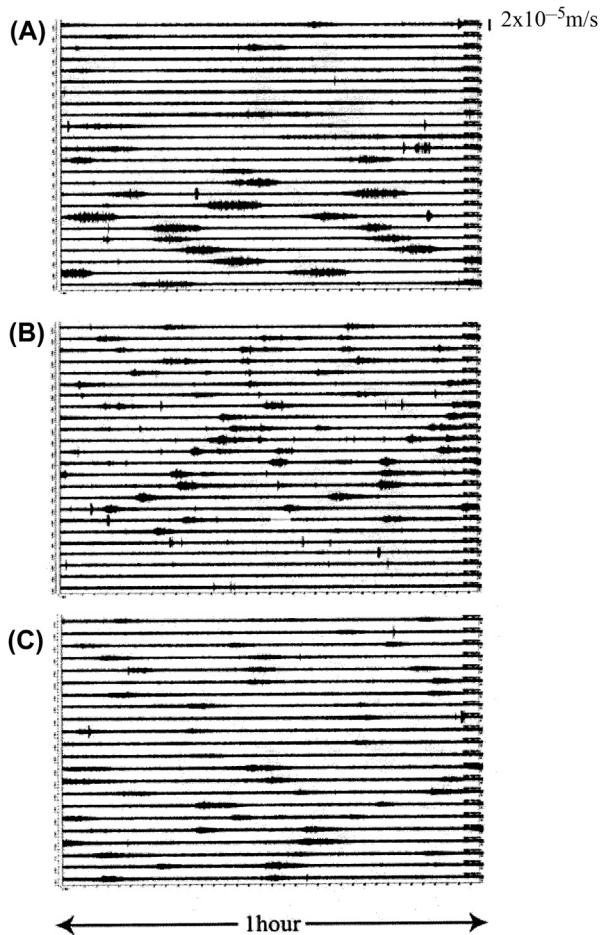


**FIGURE 12.22** Examples of waveforms of banded tremor recorded at Miyakejima volcano on (A) August 15, 2002, (B) August 11, 2003, and (C) March 29, 2004. Each plate includes a 1-day seismic record consisting of 24 traces of 1-h records of vertical components of a velocity seismometer installed at about 100 m depth at the distance of about 1.8 km from the summit. Spindle-shaped packets, called as “banded tremor,” are recorded intermittently at similar intervals. From Fujita, E., 2008. Banded tremor at Miyakejima volcano, Japan: implication for two-phase flow instability. *Journal of Geophysical Research* 113, B04207.

### 12.5.1.5 Etna Volcano

The duration of the banded tremor episodes is shown in Fig. 12.24A and B, and the intervals between them were studied at Etna volcano during its low-degassing activity occurring just after the end of the May 13, 2008 to July 6, 2008 effusive–explosive eruption from an eruptive fissure (Cannata et al., 2010). It was found that the first studied period, August 20, 2008, to September 9, 2008, was characterized by 620 clear episodes of banded



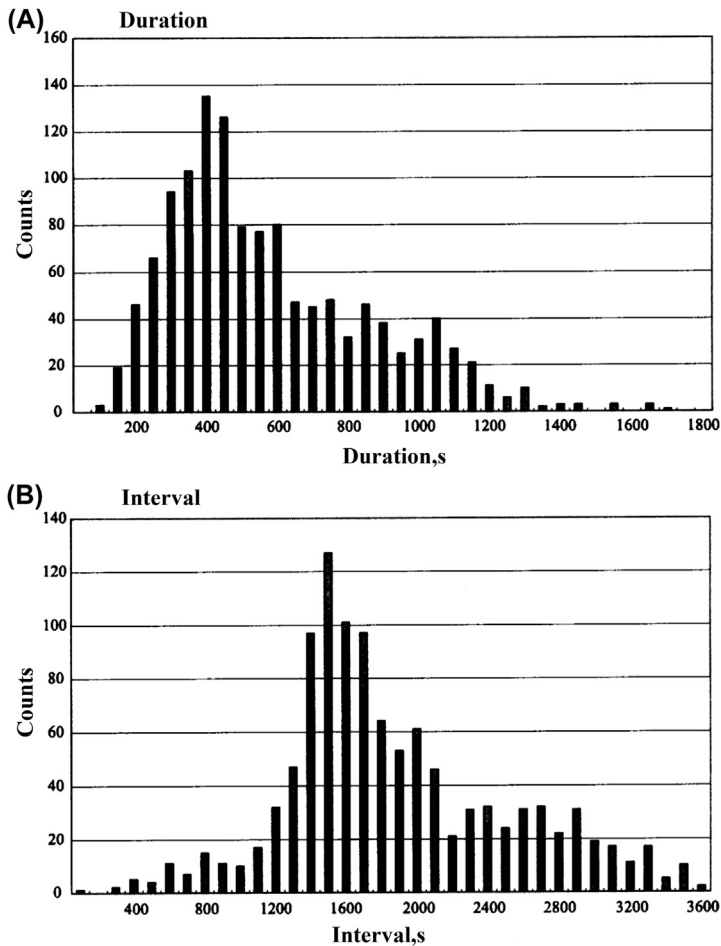


**FIGURE 12.23** The histograms of (A) duration and (B) intervals of banded tremor at Miyakejima volcano. From Fujita, E., 2008. Banded tremor at Miyakejima volcano, Japan: implication for two-phase flow instability. *Journal of Geophysical Research* 113, B04207.

tremor with fairly steady duration for the entire period and equal to about 25 minutes (Fig. 12.24C). The interval between the end of an episode and the onset of the following one was also quite steady and roughly equal to the time duration of the episodes (Fig. 12.24E). During the second period of October 1, 2008, to October 27, 2008, 650 episodes of banded tremor occurred, with an average episode duration of about 30 minutes and intervals between the end of an episode and the onset of the following one roughly equal to 25 minutes (Fig. 12.24D and F).

This type of distribution of the banded tremor episode durations and the intervals between them, obtained for Etna and Miyakejima, suggest the

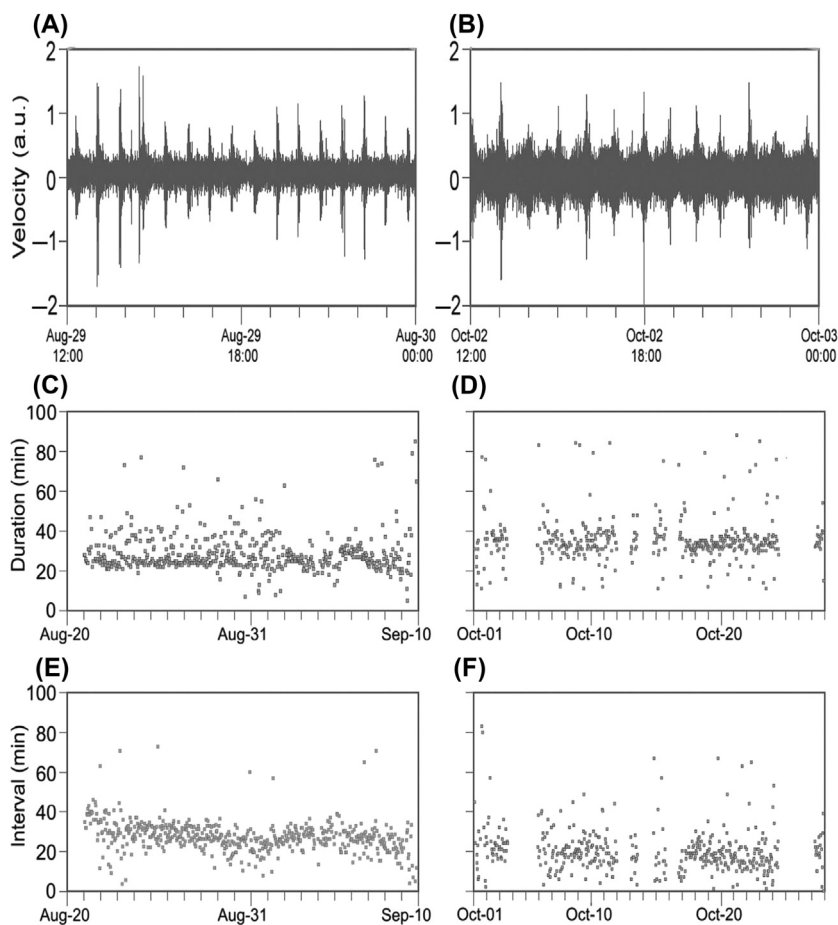


**FIGURE 12.24** Examples of broadband seismic signals of banded tremor at Etna volcano recorded at a distance of about 1 km from the active vent (A, B), the duration of observed banded tremor episodes (C, D), and the length of the intervals between the end of one episode and the start of another (E, F). *Modified from Cannata, A., Di Gracia, G., Montalvo, P., Ferrari, F., Nunnari, G., Patanè, D., Privitera, E., 2010. New insights into banded tremor from the 2008–2009 Mount Etna eruption. Journal of Geophysical Research 115, B12318.*

existence of some threshold to trigger the banded tremor (Fujita, 2008). The important factor for the understanding of the nature of banded tremor is also its absence during the eruption activity. Fujita (2008) proposed the shallow hydrothermal system as the most plausible mechanism for the generation of banded tremor. The shallow water aquifers appear to be capable of maintaining continuous gas emission and triggering oscillations while magma plays only a passive role specifically as a heat source. Fujita (2008) modeled the source of this phenomenon with a two-phase hydrothermal instability flow

model, consisting of subcool and superheated water regions. The interval between banded tremor episodes can be interpreted in terms of this two-phase flow oscillation with a heat supply from the magma beneath the hydro-thermal system.

### 12.5.2 Long-Period Tremor

Long-period tremors with periods of oscillations longer than 1 second were observed at several volcanoes of the world during different stages of eruption activity and without any eruption activity. The results of its location and source mechanism are presented in this section.

#### 12.5.2.1 Aso Volcano, Kyushu

Long-period ( $T > 7$  seconds) tremor signals of rather short duration (several tens of seconds) have been observed at Aso volcano by a broadband network in 1984 (Kawakatsu et al., 2000). These long-period signals occurred at the volcano during the small phreatic explosions or without surface activity (Fig. 12.25). The most energetic period of 15 seconds is always clearly recorded.

The events were located within the depth range from 1.2 to 1.6 km beneath the volcanic crater. Legrand et al. (2000) used a linear inversion to estimate the moment tensor of the events. The inversions showed a similar isotropic mechanism for all 43 long-period tremors and the seismic signals of six phreatic explosions.

#### 12.5.2.2 Usu, Hokkaido

Long-period volcanic tremors were observed at Usu volcano by Yamamoto et al. (2002) during the first weeks of the 2000 eruptive activity just after the March 31 and April 1 phreatic explosions (see Section 7.2). Long-period

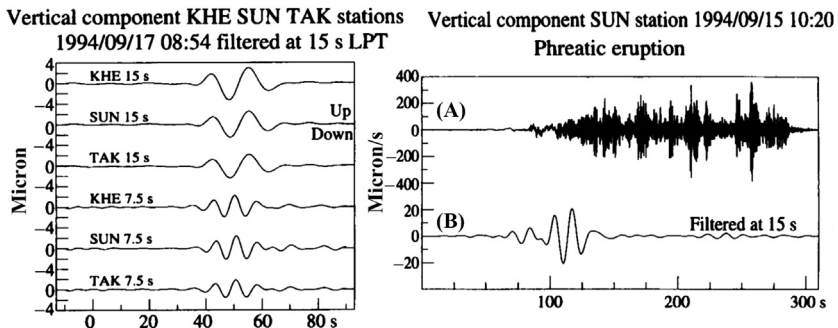


**FIGURE 12.25** Broadband records of long-period seismic events recorded at Aso volcano, Kyushu, without surface activity and during the phreatic explosion. Both types of records filtered at 15 s have comparable waveforms. From Legrand, D., Kaneshima, S., Kawakatsu, H., 2000. Moment tensor analysis of near-field broadband waveforms observed at Aso volcano, Japan. *Journal of Volcanology and Geothermal Research* 101, 155–169.

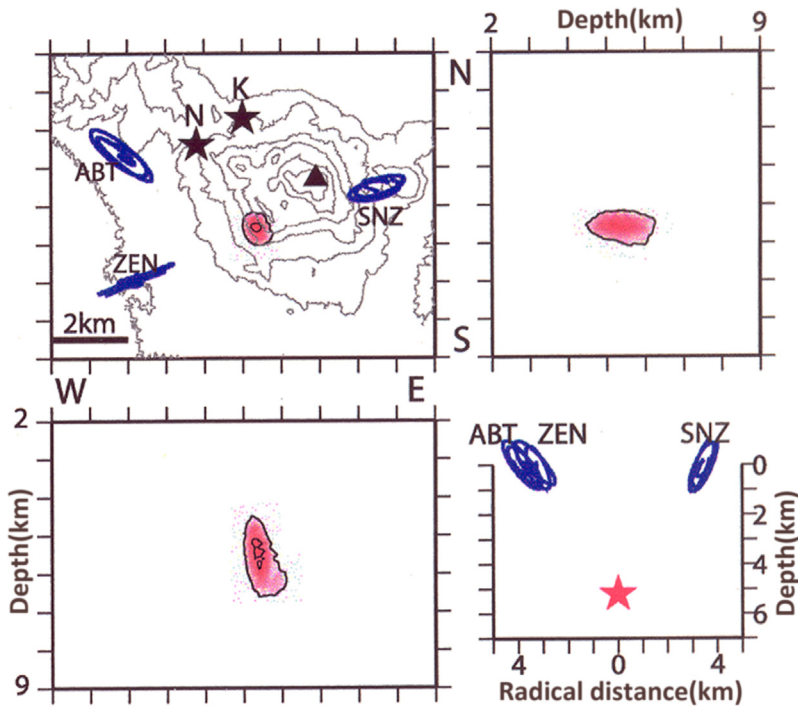


**FIGURE 12.26** The source location of the long-period tremor that was observed just after the 2000 phreatic explosions from craters N and K at Usu volcano, Japan. ABT, ZEN, and SNZ are broadband seismic stations. From Yamamoto, M., Kawakatsu, H., Yomogida, K., Koyama, J., 2002. Long-period (12 sec) volcanic tremor observed at Usu 2000 eruption: seismological detection of a deep magma plumbing system. *Geophysical Research Letters* 29. <http://doi.org/10.1020/2001GL013996>.

volcanic tremors were recorded as isolated wave packets. The period of these tremors varied gradually from 10 to 12 seconds during the first 2 or 3 days after the first eruption and became steady at 12 seconds. The source of long-period tremors was located by the waveform semblance method. The estimated source region was about 1.5 km southeast of the summit crater at a depth of 5–6 km (Fig. 12.26). The best fitting seismic moment tensor, estimated for the April 7, 2000 isolated wave packet of long-period volcanic tremor, corresponded to a spherical source and consisted of an isotropic component (the amplitude ratio  $M_{xx}:M_{yy}:M_{zz}$  is 1:1:1) and a CLVD component, with the ratio of these two components equal to (3.75):(–1).

### 12.5.2.3 Okmok Volcano, Alaska

The continuous, long-period (2.5–5 seconds) tremor was recorded during the 2008 eruption of Okmok Volcano, Alaska (Haney, 2010). The wave field of the tremor consisted mainly of surface waves, of both Rayleigh and Love types.

The procedure based on the interstation arrival times computed with cross-correlation yields the source location at the depth of 2 km close to the new cone built over the course of the eruption. Such a depth places the tremor source between the caldera floor and the shallow ( $> 3$  km b.s.l.) magma reservoir at Okmok.

### 12.5.3 Deep Tremor

Deep tremors may be distinguished from shallow ones by their amplitude distribution. For Kilauea, Hawaii, deep tremor shows more uniform amplitude for many stations over a large area as compared to shallow events, which show the concentration of large amplitudes only at a few nearby stations. Deep Hawaiian tremor demonstrated an identical period at almost all stations but it had a tendency to change with time. Fig. 12.27 shows the source locations of 24 deep

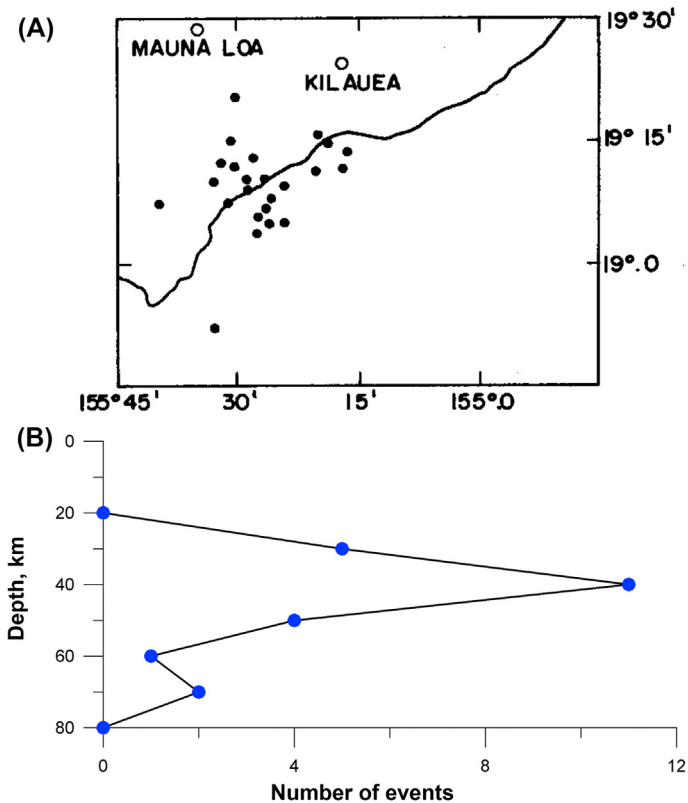


**FIGURE 12.27** Locations of the 1973–77 deep tremor sources under Kilauea volcano, Hawaii (A) and their depth distribution (B). From Aki, K., Koyanagi, R., 1981. Deep volcanic tremor and magma ascent mechanism under Kilauea, Hawaii. *Journal of Geophysical Research* 86, 7095–7109.

tremor episodes that were determined using the onset times. The tremor sources were distributed at a distance from 13 to 40 km to the southwest of Kilauea volcano within a depth range of 27–75 km (maximum number of events were recorded between the depths from 35 to 45 km) with the mean error in-depth estimation of about 6 km (Aki and Koyanagi, 1981).

Aki and Koyanagi (1981) noted that the deep magma transport inferred from the deep tremor data is relatively unaffected by volcanic activity near the surface. At the same time, numerous publications, that appeared beginning in 2002 (Obara, 2002; Nadeau and Dolenc, 2005; Payero et al., 2008), show the existence of deep (between 20 and 40 km of depth) nonvolcanic tremors on some subduction zone thrust faults and on San Andreas fault. A number of studies associate this deep tremor as a manifestation of the process of the transition zone between the seismogenic coupled and deep free-slipping segments of the subduction interface. Therefore it is possible that the deep tremor beneath Hawaii is also of the same nature but not of volcanic origin.

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## **RELEVANT WEBSITE**

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## Chapter 13

# Seismic Signals Associated With Pyroclastic Flows, Rockfalls, and Lahars

Volcanic hazard is significantly related to the propagation of density currents such as pyroclastic flows and lahars. These flows represent the motion of natural sediments generated along volcanic slopes that propagate with high velocity and can devastate everything in their path. Both types of volcanic flows have produced destructive effects during human history. In the list of the 27 largest volcanic catastrophes between 1700 and 1986 that were responsible for at least 1000 deaths (Bardintzeff and McBirney, 1999), 11 were associated with pyroclastic flows and 11 with lahars. Therefore the ability to monitor and track these phenomena in real time is important for the mitigation of volcanic risk.

### 13.1 OCCURRENCE OF PYROCLASTIC FLOWS, ROCKFALLS, AND LAHARS DURING VOLCANIC ERUPTIONS

Fig. 13.1 shows the typical occurrence of pyroclastic flows, rockfalls, and lahars during the 2004–06 effusive–explosive volcanic activity at Volcán de Colima, Mexico. The extrusion of lava generated numerous pyroclastic flows and rockfalls in October 2004 due to partial collapses of the summit dome. The next, and more voluminous, portion of pyroclastic flows was produced by the fall of volcanic column material associated with large explosions in 2005. Lahars were observed during the rainy seasons (from June to October) in 2005 and 2006. Their number was larger in 2005 just after lava extrusion and the large explosions. In 2006 their number decreased with decreasing eruptive activity.

### 13.2 SEISMIC SIGNALS ASSOCIATED WITH PYROCLASTIC FLOWS AND ROCKFALLS: WAVEFORMS AND SPECTRA

Pyroclastic flows and rockfalls are common for both explosive and effusive activity at andesitic and dacitic volcanoes. They may be generated by a

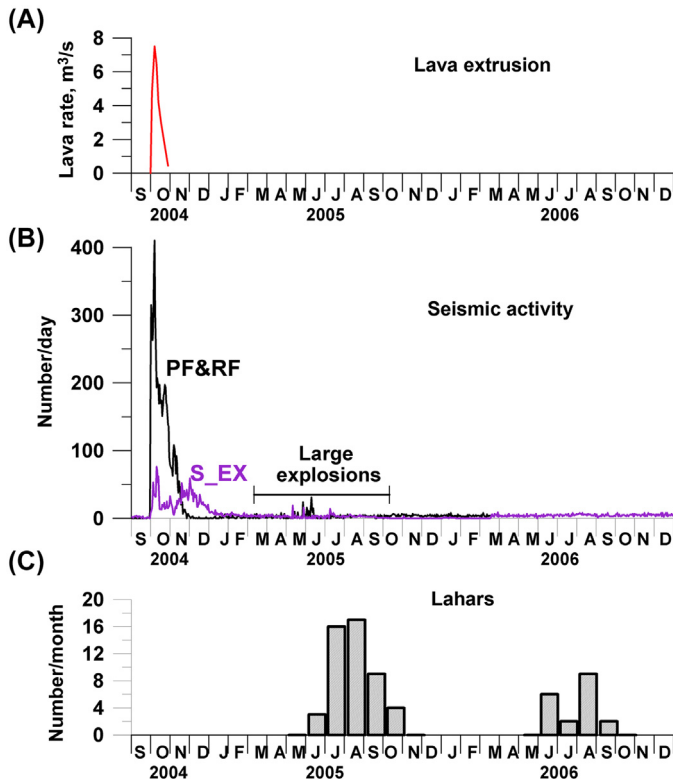


**FIGURE 13.1** Sequence of volcanic events during the 2004–05 effusive–explosive episodes at Volcán de Colima. (A) Variations in the rate of lava extrusion (Zobin et al., 2006b). (B) Seismic activity associated with the occurrence of pyroclastic flows and rockfalls (PF&RF) and with small gas-and-ash explosions (S\_EX). The period of large explosions is shown. (C) The occurrence of lahars. From Zobin, V.M., Plascencia, I., Reyes, G., Navarro, C., 2009a. The characteristics of seismic signals produced by lahars and pyroclastic flows: Volcán de Colima, México. *Journal of Volcanology and Geothermal Research* 179, 157–167.

number of different mechanisms. Among them are the partial gravitational collapse of a growing lava dome, the collapse from an eruption column, and the explosive destruction of a growing lava dome (Francis, 1998; Carey and Bursik, 2000). As noted by Calder et al. (2002), pyroclastic flows are distinguished from rockfalls by their larger size, longer run out ( $\geq 0.5$  km), greater production of fine ash, and appreciable buoyant hot ash clouds. The fall and flow of fragments of lava domes or flow fronts and the numerous rockfalls generate seismic signals.

Pyroclastic flows and rockfalls are associated with similar emergent, cigar-shaped waxing, and waning seismic signals with frequencies of

1–10 Hz, which differ only in their durations and/or amplitudes (Calder et al., 2002). The signals produced by pyroclastic flows and rockfalls may represent the main component of seismicity during block-lava emission, subsequent lava-dome destruction, and explosive activity.

### 13.2.1 Seismic Signals of Pyroclastic Flows Produced by the Partial Collapse of Lava Dome

Fig. 13.2A and B shows a broadband seismic record of pyroclastic flow associated with the partial lava-dome collapse occurring at Soufrière Hills volcano on January 8, 2007, and recorded at a distance of about 2 km from the crater. De Angelis et al. (2007) give the detailed characteristics of the observed time series. These signals can be divided into three main phases. Phase 1 is characterized by a steady increase in amplitude with time and is supposed to represent the progressive failure of slip zones with variable orientation along the boundaries of an unstable portion of the dome as the result of increased internal pressurization. Phase 2 consists of a strong spindle-shaped pulse; the onset is emergent and no clearly defined seismic phases could be recognized in the seismic records. This phase is thought to represent the fall of the unstable portion of the lava dome, its collision with the slopes of the volcano and resultant fragmentation, and the descent of the flows down the flanks of the edifice. Phase 3 includes a sequence of seismic pulses possibly due to the remobilization of material deposited along the same path by the pyroclastic flow during phase 2.

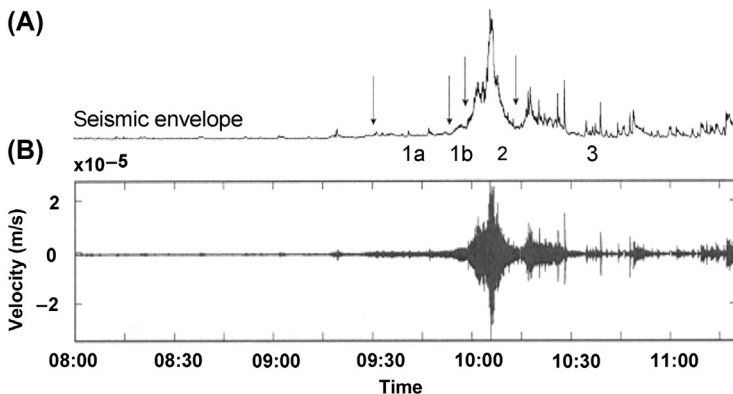


**FIGURE 13.2** Seismic envelope (A) and broadband velocity seismogram (vertical component) (B) for the signals associated with pyroclastic flow at Soufrière Hills volcano, recorded on January 8, 2007, at a distance of about 2 km from the crater. The seismogram is divided into different phases according to variable observed activity (see text). After De Angelis, S., Bass, V., Hards, V., Ryan, G., 2007. *Seismic characterization of pyroclastic flow activity at Soufrière Hills volcano, Montserrat, 8 January 2007. Natural Hazards and Earth System Sciences* 7, 467–472.

Fig. 13.3 shows broadband seismic records associated with the pyroclastic flows produced by the partial collapse of a lava dome during the 2004 lava extrusion at Volcán de Colima, Mexico and recorded at a distance of 4 km from the crater. They correspond to phase 2 in Fig. 13.2. The waveforms begin with a low-amplitude signal which gradually increases in amplitude and then more slowly decreases. Fig. 13.4 shows the details of the initial part of the record. It consists of an initial relatively short-period impulse (1) of duration of about 2 seconds, followed by a long-period

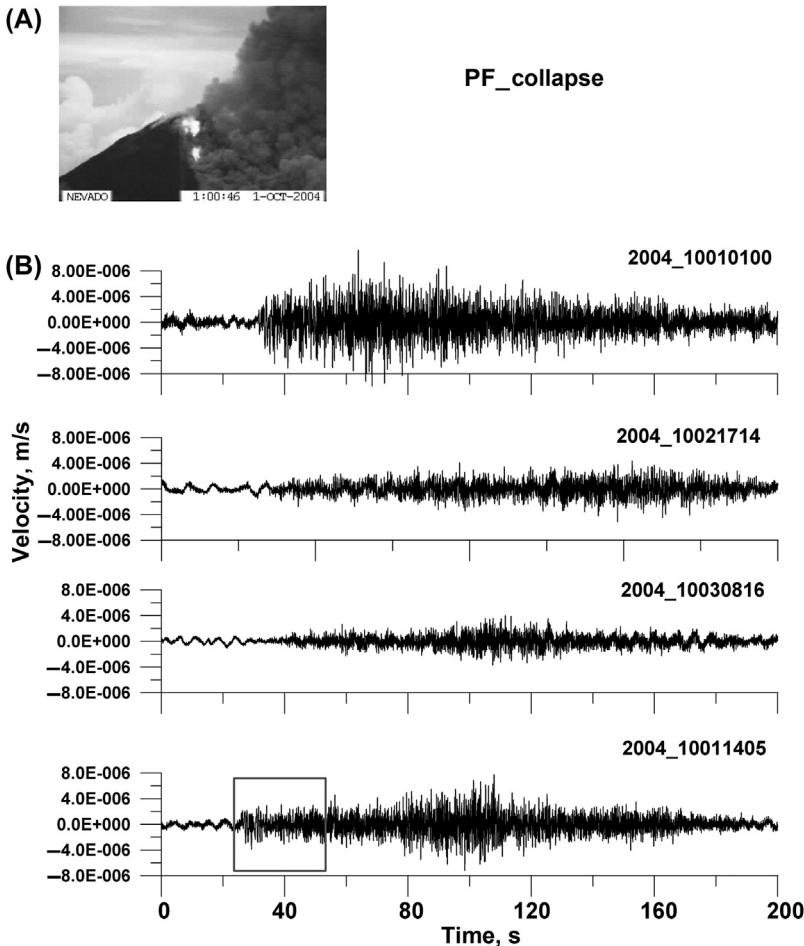


**FIGURE 13.3** Video image of the event 2004\_10010100 (A) and broadband seismic records (vertical component) of pyroclastic flows produced by the partial collapse of the lava dome at Volcán de Colima and recorded at a distance of 4 km from the crater (B). The index near the seismic record represents the date (yyyy\_mmddhhmm) of the events. The rectangle on the seismogram of 2004\_10011405 shows the selected sample for Fig. 13.4. Courtesy Colima Volcano Observatory.

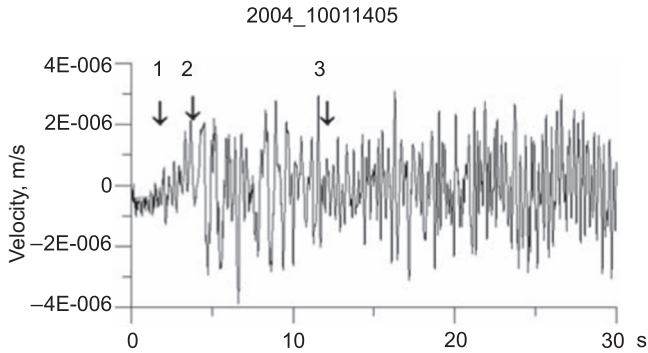


**FIGURE 13.4** Details of seismic record 2004\_10011405 of the pyroclastic flow that was shown in Fig. 13.3 in rectangle. Arrows with numbers show different seismic impulses (see text). Courtesy Colima Volcano Observatory.

impulse (2) of duration of about 9 seconds, and then a long-duration short-period impulse (3).

Comparison with the seismograms and video images observed for the Unzen volcano pyroclastic flows (Yamasato, 1997; see Chapter 11.) showed that impulse 1 may be produced by the lava-dome collapse, impulse 2 corresponds to the downfall of the lava blocks onto the slope, and impulse 3 is associated with the generation of pyroclastic flows. The durations of the first two impulses are comparable with those reported for the Unzen pyroclastic flows (Yamasato, 1997).

### 13.2.2 Seismic Signals of Pyroclastic Flows Produced by the Collapse of Eruption Column

Fig. 13.5 shows broadband seismic records associated with the pyroclastic flows emplaced from the eruption columns during the four large 2005 Vulcanian explosions at Volcán de Colima, Mexico. Pyroclastic flows entered the Colima ravines and reached distances up to 3–6 km from the summit. Satellite imageries identified the ash clouds produced during these eruptions at heights between 7.6 and 8.5 km above sea level (a.s.l.). The seismic station was situated at a distance of 4 km from the crater.

The records of this type of pyroclastic flows are more complex compared with those emplaced from the partial collapse of the dome. The detailed structure of the seismic record “explosion – pyroclastic flow” is shown in Fig. 13.6. The seismic signal begins with the large amplitude records of the explosions; then the collapse of volcanic material from an eruption column generates a pyroclastic surge. The generation of pyroclastic flow begins 20–25 seconds after the onset of the seismic signal. The record of an explosion has significantly larger amplitudes than the signal of a

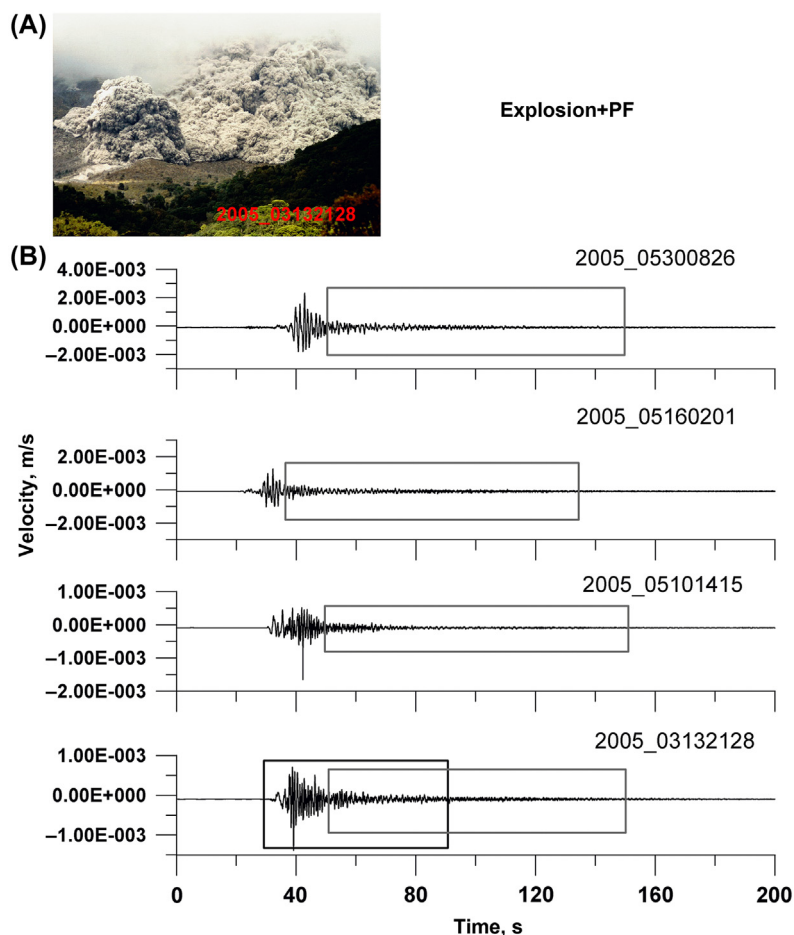


**FIGURE 13.5** Photo of pyroclastic flow emplaced from eruption column of the 2005\_03132128 explosion at Volcán de Colima (A) and broadband seismic records (vertical component) of explosions and following pyroclastic flows emplaced from eruption columns at Volcán de Colima during its 2005 activity and recorded at a distance of 4 km from the crater (B). The bold line rectangle on the seismogram of 2004\_10011405 shows the selected sample for Fig. 13.6. The thin line rectangles show the selected samples for Fig. 13.7. Courtesy Colima Volcano Observatory. Photo of the 2005\_03132128 explosion was taken by Sergio Velasco.

pyroclastic flow. We can identify at least five impulses. As was proposed by [Zobin et al. \(2006\)](#), the first two impulses are generated by the movement of fragmented magma above the fragmentation surface in the shallow conduit (1) and by an explosion in the conduit (2). Impulses 3–5 are equivalent to impulses 1–3 in Fig. 13.3 for the pyroclastic flows originating from a lava-dome collapse: impulse 3 is generated by the rockfall from the eruptive

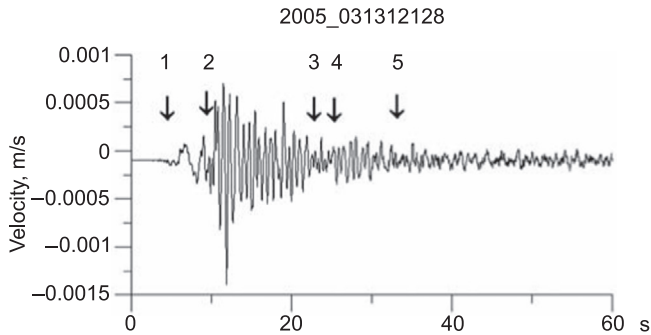


**FIGURE 13.6** A structure of the seismic record of explosion 2004\_10011405 at Volcán de Colima and the following pyroclastic flow shown in Fig. 13.5. Arrows with number show the arrivals of the seismic impulses (see text). *Courtesy Colima Volcano Observatory.*

column, impulse 4 corresponds to the movement of the rocks along the slope, and impulse 5 is generated by the pyroclastic flow. The waveform of impulse 5 is more heterogeneous than that of impulse 3 in Fig. 13.3. The fall of volcanic rocks from the eruptive column and generation of pyroclastic flow begins, therefore, 20–25 seconds after the onset of the seismic signal. The continuous fall of the rocks from the column and their following movement adds the low-frequency component in this phase.

Fig. 13.7 shows a 100-second portion of the seismic records of Fig. 13.5 following the explosive onset. To remove the low-frequency noise, they were high-pass filtered at 2.5 Hz. After filtering, they are similar enough to the records of pyroclastic flows generated by the collapse of a lava dome. At the same time, their amplitudes are one order higher than for the seismic records of the pyroclastic flows initiated by the partial collapse of a lava dome.

### 13.2.3 Seismic Signals of Pyroclastic Flows Produced by the Explosive Destruction of Growing Lava Dome

Fig. 13.8B shows a short-period seismic record associated with the May 9, 2006 explosive eruption and following pyroclastic flow at the andesitic Bezymianny volcano, Kamchatka. The Vulcanian explosion occurred during the stage of the lava-dome growth in the crater that was observed from the end of April 2006. An explosive eruption produced the ash column that rose to a height of  $\sim 15$  km a.s.l (Fig. 13.8). Pyroclastic flow, generated due to the partial explosive destruction of the growing lava dome, extended 7–8 km from the crater. Video monitoring of this eruption (Droznin and Droznin, 2007) indicated that the emplacement of the pyroclastic flow began about 5–7 minutes after the explosion.

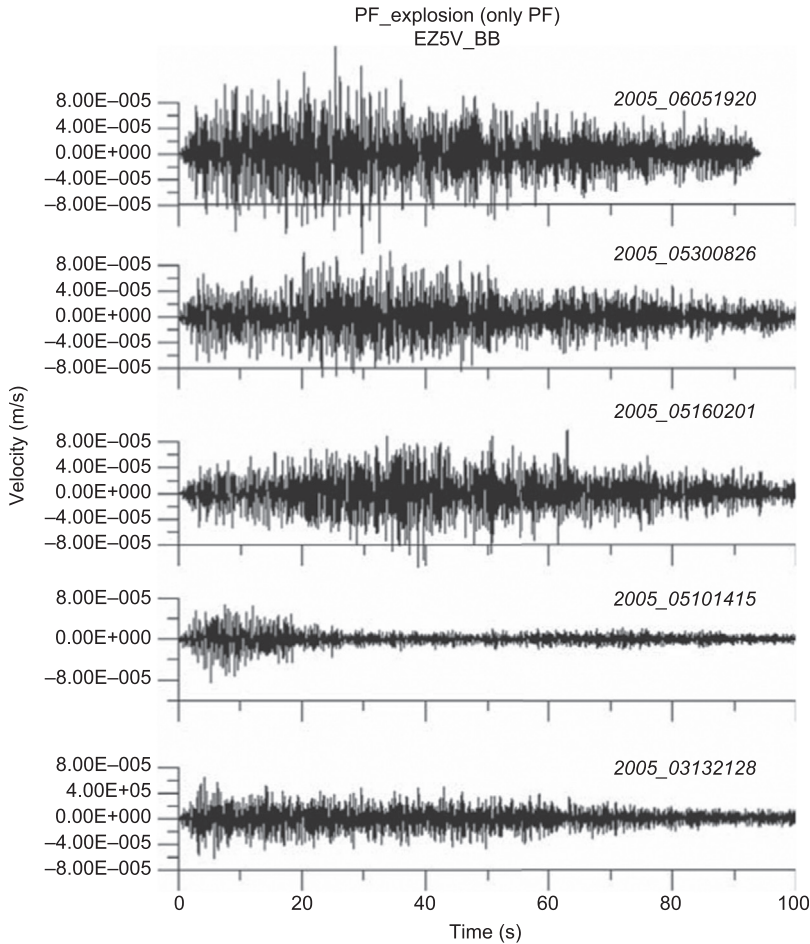


**FIGURE 13.7** Seismic records (samples in rectangles selected in Fig. 13.5) of pyroclastic flows following explosions and emplaced from the eruption columns at Volcán de Colima. *Courtesy Colima Volcano Observatory.*

The seismic record, shown in Fig. 13.8B, illustrates this sequence of eruptive events. The initial signal corresponding to the Vulcanian explosion (shown in ellipse) is separated from the secondary group of seismic signals associated with the propagation of pyroclastic flow (shown in a rectangle) with a time interval of about 6–7 minutes as was noted by the video monitoring. The duration of the seismic signal, associated with a pyroclastic flow, is about 16 minutes. The spindle-shaped waveform of the signal is generally similar to the waveforms produced by pyroclastic flows generated by the partial collapse of the lava dome and the collapse of the eruption column.



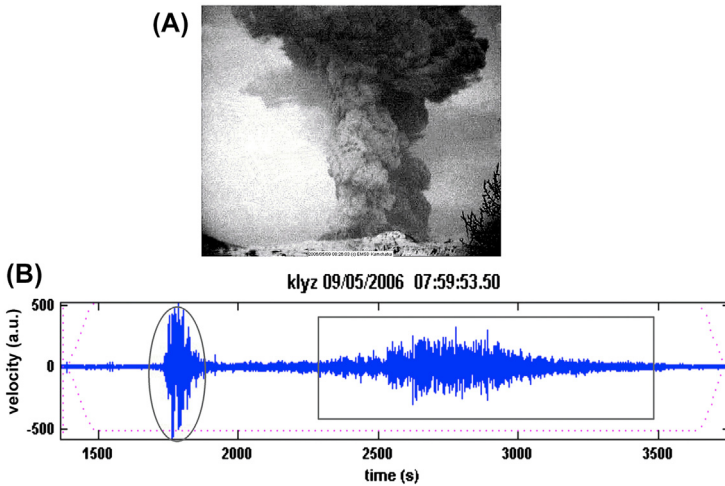


**FIGURE 13.8** Video image of the May 9, 2006 eruption column at Bezmianny volcano (A) and short-period seismic record (velocity, horizontal component) associated with the Vulcanian explosion (shown in ellipse) and the following pyroclastic flow (shown in rectangle) (B). The seismic signal was recorded at a distance of 40 km from the crater. AU means arbitrary units. *Seismic record is courtesy of the Kamchatkan Methodical Seismological Department of the RAS Geophysical Survey, Russia. Video image is taken from Droznin, V.A., Droznin, D.V., 2007. The 9 May 2006 activity of Bezmianny volcano. Proc. Kamch. Region. Sci. Cent., Earth Sci. vol. 9, 105–110 (in Russian).*

#### 13.2.4 Seismic Signals Produced by Rockfall

Fig. 13.9 shows the seismic records associated with the rockfalls that were observed during the 2004–05 effusive–explosive episodes at Volcán de Colima, Mexico. The waveforms are similar to those produced by pyroclastic flows although shorter. The same observation emerges from the comparison of the seismic signals produced by rockfalls and pyroclastic flows at Soufrière Hills volcano (De Angelis et al., 2007; Fig. 13.10).

#### 13.2.5 Spectral Characteristics

Fourier spectra of the seismic signals, associated with the emplacement of pyroclastic flows and recorded at a distance of 4 km from the crater of Volcán de Colima, are shown in Fig. 13.11A. They were calculated for broadband seismograms including the seismic records of pyroclastic flows initiated by a partial lava-dome collapse (PF\_collapse) and the seismic records of an explosion and following pyroclastic flow (Explosion + PF) and of the pyroclastic flow only (PF\_explosion). The spectra of both types of pyroclastic flows are similar with a close range of peak frequencies (between 2.4 and 4.4 Hz).

The Fourier spectrum of the seismic signal, produced by the pyroclastic flow that followed the May 9, 2006 explosive destruction of the lava dome

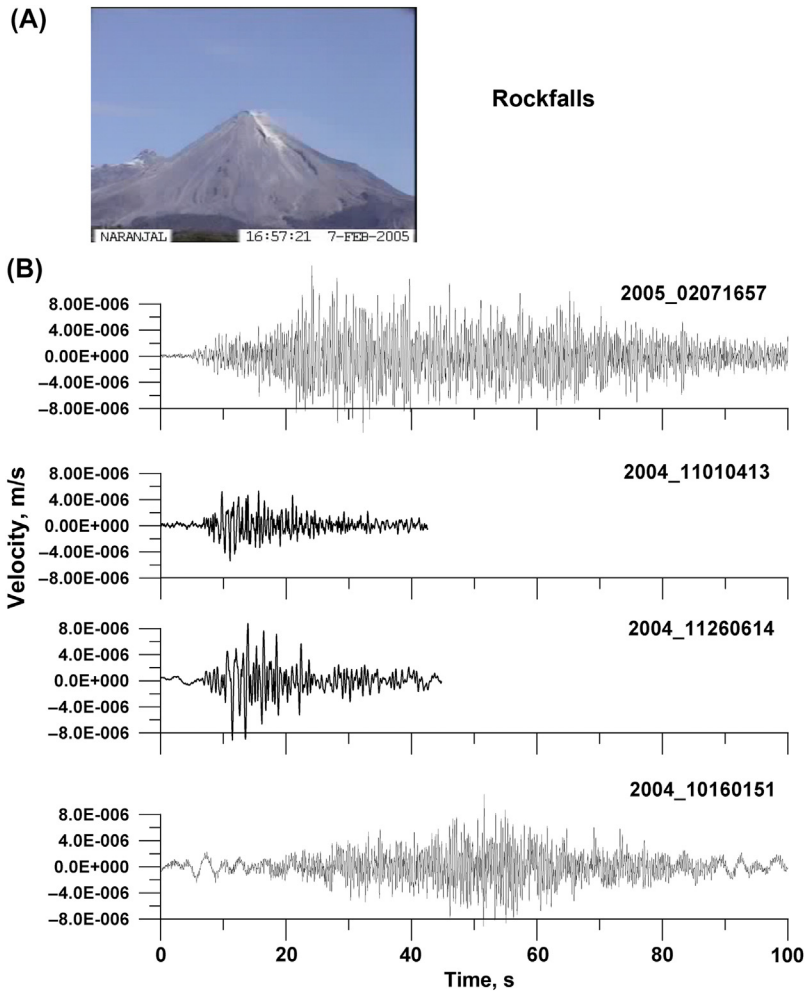


**FIGURE 13.9** Video image of rockfall corresponding to the event 2005\_02071657 (A) and broadband seismic records (vertical component) of rockfalls observed at Volcán de Colima during the 2004–05 effusive–explosive episodes and recorded at a distance of 4 km from the crater (B). *Courtesy Colima Volcano Observatory.*

at Bezymianny volcano and was recorded at a distance of 40 km (Fig. 13.11B), had its maximum amplitude at 0.7 Hz, lower than for the seismic signals associated with the two types of pyroclastic flows mentioned previously. The duration of the seismic signal at Bezymianny was significantly longer than the signals recorded during the emplacement of the pyroclastic flows observed at Volcán de Colima which may be a reason for the generation of lower frequencies of the seismic record.

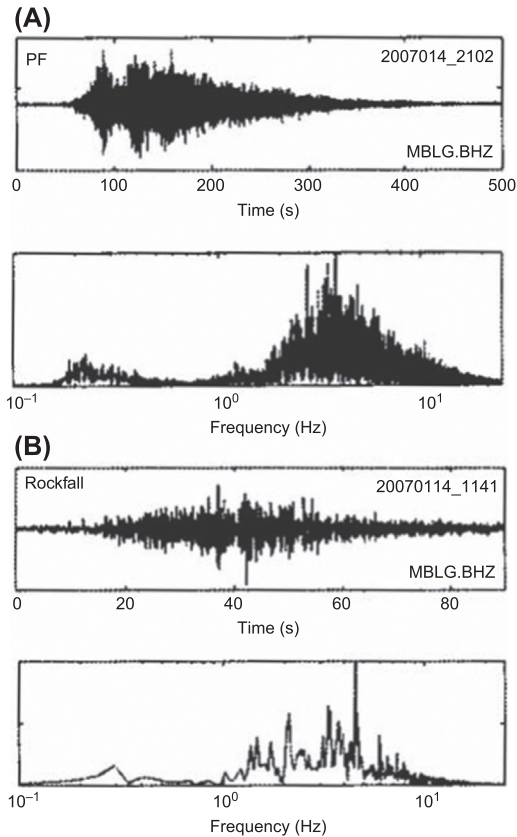


**FIGURE 13.10** Short-period seismic records associated with the pyroclastic flow (A) and rockfall (B) observed during the 2007 activity at Soufrière Hills volcano and Fourier spectra of these signals. After De Angelis, S., Bass, V., Hards, V., Ryan, G., 2007. *Seismic characterization of pyroclastic flow activity at Soufrière Hills volcano, Montserrat, 8 January 2007. Natural Hazards and Earth System Sciences* 7, 467–472.

Spectra of the seismic signals associated with rockfalls occurring at Volcán de Colima (Fig. 13.11A) are characterized by a tendency to be of lower frequency for shorter seismic records and their spectral peaks vary between 1 and 3 Hz.

### 13.3 OCCURRENCES OF EARTHQUAKES ASSOCIATED WITH PYROCLASTIC FLOWS AND ROCKFALLS

The earthquakes associated with pyroclastic flows and rockfalls occur in dense groups, sometimes completely filling the seismogram (see Image 2.13 from Chapter 2: Seismicity at Volcanoes). The total number of these events recorded

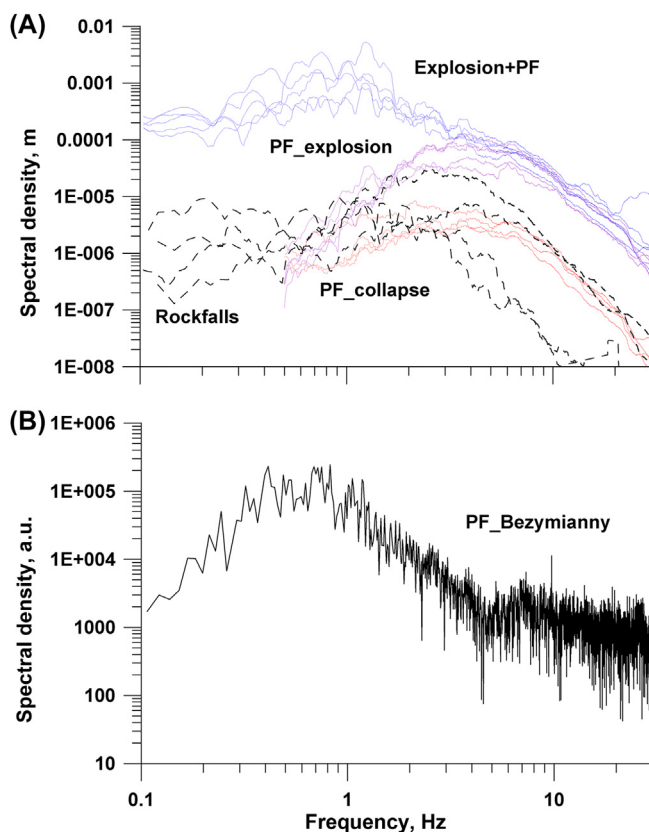


**FIGURE 13.11** Spectra of broadband seismic signals associated with pyroclastic flows and rockfalls at Volcán de Colima corresponding to those shown in Figs. 13.3, 13.5, 13.7, and 13.9 (A) and the spectrum of the short-period seismic signal associated with Vulcanian explosion and following pyroclastic flow at Bezymianny volcano shown in Fig. 13.8 (B). In (A), heavy red lines show spectra of the signals associated with the pyroclastic flows produced by the partial collapse of lava dome (PF\_collapse); thin blue lines show spectra of the signals including the seismic record of the explosion and of the following pyroclastic flow emplaced from the volcanic column during the explosion (Explosion + PF); heavy violet lines (light gray lines in print versions) show spectra of the part of these signals associated with pyroclastic flows only (PF\_explosion); and dashed black lines show spectra of the signals associated with rockfalls (Rockfalls). *Courtesy Colima Volcano Observatory.*

during the 80-day andesitic block-lava extrusion at Volcán de Colima, Mexico in November 1998–February 1999 exceeded 6000 (Zobin et al., 2005). Pyroclastic flow and rockfall signals replaced the volcano-tectonic earthquakes from the preceding swarm just after the start of lava extrusion. The rate of seismic energy release during the seismic activity of pyroclastic flows and rockfalls was comparable with the rate of seismic energy release during the precursory swarm of volcano-tectonic earthquakes (Fig. 13.12).

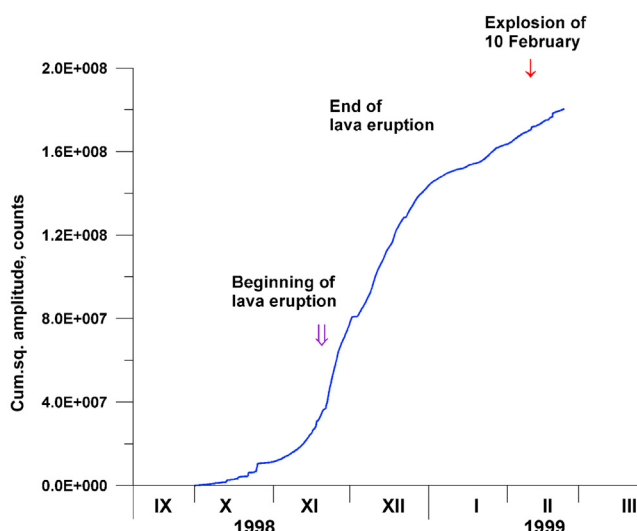


**FIGURE 13.12** Cumulative real-time seismic energy measurement (RSEM) curve of variations in seismic energy release at Volcán de Colima from October 1998 to February 1999. It is seen that the rate of seismic energy release, generated by volcano-tectonic earthquakes and recorded before the eruption, did not significantly change with the beginning of the lava eruption when the seismic signals were generated mainly by pyroclastic flows and rockfalls. *From Zobin, V.M., Orozco-Rojas, J., Reyes-Dávila, G.A., Navarro, C., 2005b. Seismicity of an andesitic volcano during block-lava efusión: Volcán de Colima, México, November 1998–January 1999. Bulletin of Volcanology 67, 679–688.*

Malone (1983) reported that the typical 1980–81 dome-building eruption episodes at Mount St. Helens, Cascades began with a few volcano-tectonic earthquakes several weeks before the actual eruption. At the onset of the eruption, the volcano-tectonic events are replaced by avalanche signals and explosion events. Both types of events may last for several days, slowly decreasing in number.

However, pyroclastic flow and rockfall seismicity may begin without any preliminary volcano-tectonic activity as was observed, for example, during the 2002, 2004, and 2007 extrusions at Volcán de Colima (Zobin et al., 2008).

### 13.4 RELATIONSHIP BETWEEN THE PYROCLASTIC FLOW AND ROCKFALL EARTHQUAKES AND SEISMOVOLCANIC ACTIVITY DURING THE LAVA EXTRUSION

The relationship between the seismic signals, produced by pyroclastic flows and rockfalls, and eruption dynamics has been discussed by several authors. Neuberg (2000) reported that these signals correlated visually with the avalanches of material from the lava dome at Soufrière Hills volcano.

Ratdomopurbo and Poupinet (2000) noted that the duration of the seismic signals (“guguran” in Indonesian) at Merapi volcano, Java is several minutes. Their frequencies were higher than that of tremors. Purbawinata et al. (1997) suggested that the occurrences of “guguran” may indicate the level of stability of the lava dome. When the dome is unstable, “guguran” occurrence may be greater than 500 events/day. Whereas when the dome is stable, only about 3 events/day occur. A rate of more than 50 events/day is usually considered as an indication that the dome is in an unstable state.

A dacitic lava dome at Unzen volcano, Japan, emerged in May 1991, and its collapse resulted in the generation of pyroclastic flows. Yamasato (1997) studied the seismic signals produced by the collapse of lava blocks from the

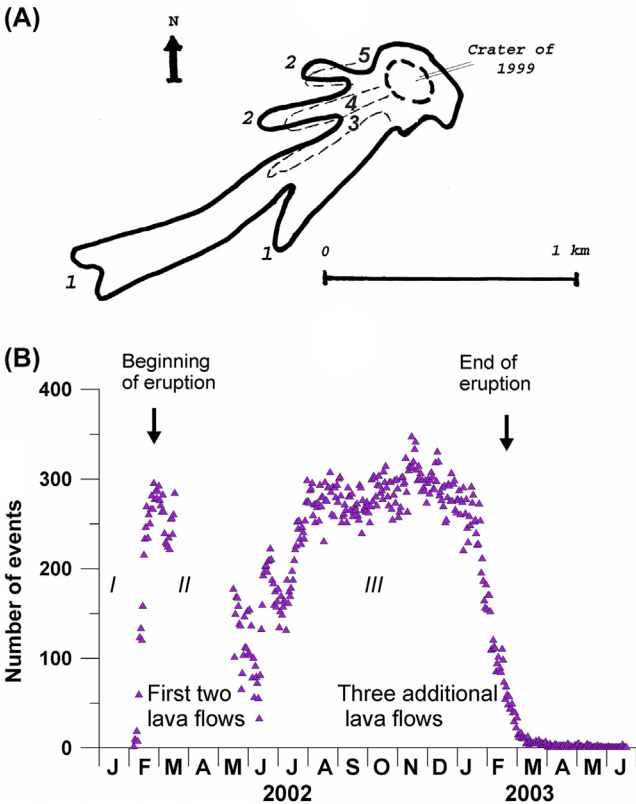


**FIGURE 13.13** Sketch of lava flows emplaced during the 2002–03 lava extrusion at Volcán de Colima (A) and the variations in the number of seismic events produced by rockfalls during January 2002 to June 2003 (B). In (A), bold dashed line outlines the crater formed during the 1999 activity. 1–5 is the numbers of lava flows according to their appearance. In (B), I–III indicates the stages of eruption described in the text. The sketch was prepared by C. Navarro, as well as the count of events. *Courtesy Colima Volcano Observatory.*

dome, the fall of material along the slope, and the migration of pyroclastic flows along the flanks of the edifice. Comparison of the seismic signals with video footage allowed the identification of seismic signals associated with the different stages of the formation of a pyroclastic flow. Small amplitude seismic waves with a predominant frequency of 2–3 Hz were excited almost simultaneously with the collapse of the lava dome. Within a few seconds, when lava blocks fell onto the slope, the amplitude of the seismic signal became large with the appearance of a low-frequency (about 0.5 Hz) component. Then the fragmented pyroclastics started to flow generating high-frequency (greater than 2 Hz) seismic waves.

Systematic variations in the number of rockfalls estimated from seismic data may be used in the monitoring of eruptive activity. Fig. 13.13 shows variations in the number of rockfalls during the 2002–03 effusive eruption at Volcán de Colima. The lava dome filled the crater up until early February 2002, and two new lava flows began their emplacements on February 14 (Fig. 13.13A). The formation of lava flows was preceded and accompanied by rockfall activity starting on the evening of February 4 and peaking on February 24 (stage I). At the end of February effusive activity significantly decreased, as reflected by the sharp decrease in the number of rockfalls in Fig. 13.13 (stage II). In May 2002, a new stage of effusive activity began, and three additional lava flows (Fig. 13.13A) were formed during June 2002–December 2002 and evolved until February 2003 (stage III). This stage corresponds to a sharp increase in the number of rockfalls until November 2002 followed by a gradual decrease in their number until March 2003.

### 13.5 QUANTIFICATION OF PYROCLASTIC FLOW AND ROCKFALL EARTHQUAKES

Seismic signals of pyroclastic flows and rockfalls can be characterized by their amplitude and duration. Fig. 13.14 illustrates the seismic signals recorded by a short-period seismograph at Volcán de Colima, Mexico. Their durations and their amplitudes vary from event to event. Uhira and Takeo (1994) demonstrated that the long-period amplitude of the seismic signals, produced by pyroclastic flows and rockfalls, may vary by about a factor of 10. Simultaneous observations with short- and long-period seismographs at Unzen volcano, Japan, showed that long-period records of the pyroclastic flow and rockfall earthquakes are significantly shorter than short-period records of the same signals (Uhira and Takeo, 1994). On the other hand, the durations of broadband and short-period seismic records, produced by pyroclastic flows and rockfalls at Volcán de Colima occurring due to the 1998 partial collapses of the lava dome, were well correlated (Fig. 13.15).

Norris (1994) reported that rockfall sequences (Mount St. Helens, Cascades and Makaopuhi crater, Hawaii) show a poor correlation between

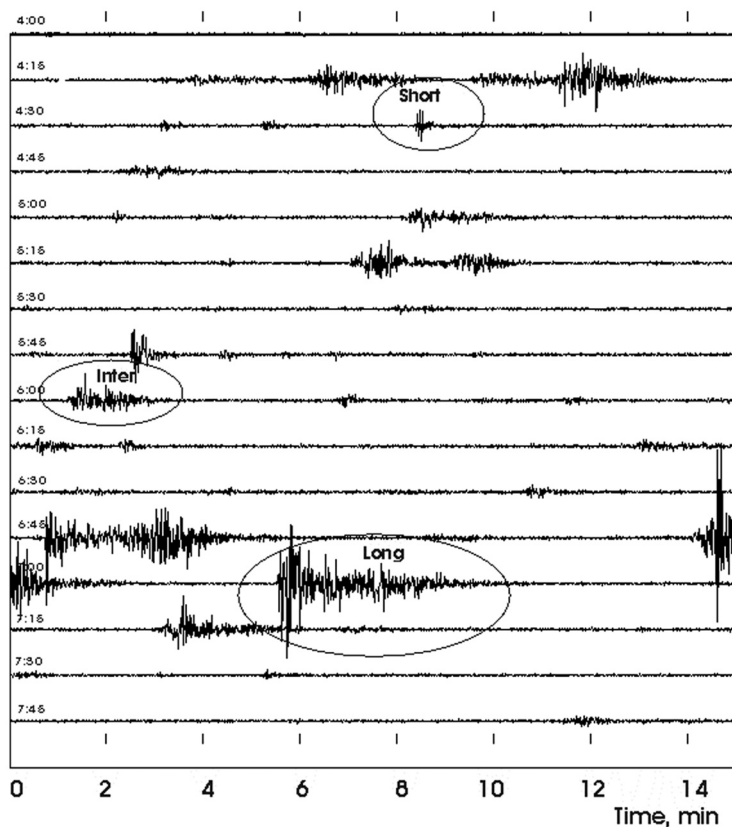


**FIGURE 13.14** Short-period seismic signals (vertical component) corresponding to rockfalls and pyroclastic flows recorded during the 4-h episode of the November 1998 lava eruption at Volcán de Colima, Mexico at a distance of 4 km from the crater. Short-, intermediate- (inter), and long-duration records are illustrated. *Courtesy Colima Volcano Observatory.*

the maximum signal amplitude recorded by a short-period seismometer and the volume of the rockfalls. [Mills \(1991\)](#) used the real-time seismic amplitude measurement (RSAM) to investigate rockfall activity at Mount St. Helens and reported a similar poor correlation; rockfalls that differed in volume by several orders of magnitude could be distinguished by their seismic amplitude or duration, but those closer in volume could not.

Comparison of the seismic data with the visual observations of pyroclastic flows at Soufrière Hills volcano, Montserrat showed that short, sharp RSAM peaks represented discrete pulses of large-volume, energetic flows ([Calder et al., 2002](#)). [Uhira and Takeo \(1994\)](#) showed a good correlation between the maximum signal amplitude recorded by long-period seismometers and the volume of pyroclastic flows at Unzen volcano,



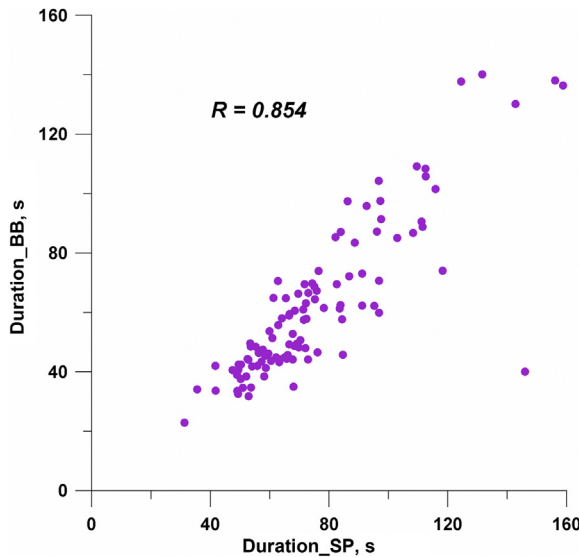


**FIGURE 13.15** Correlation between the durations of broadband and short-period seismic records of pyroclastic flows and rockfalls observed during the 2004–05 effusive–explosive episodes at Volcán de Colima.  $R$  is the coefficient of correlation. *Courtesy Colima Volcano Observatory.*

Japan. As noted by [Calder et al. \(2002\)](#), the relationship between the volume of pyroclastic flow and signal amplitude is not simple.

### 13.5.1 Quantification of Pyroclastic Flow and Rockfall Earthquakes Occurring due to Partial Collapse of the Lava Dome and Recorded by Short-Period Instruments at Volcán De Colima, Mexico

The amplitude of a short-period seismic signal is technically a less reliable parameter of an earthquake than its duration; its record may be saturated for large events. It was a reason to apply the coda magnitude  $M_c$ , based on the duration of the seismic signal, in seismic routine (e.g., [Bakun, 1984](#)). Therefore the duration of seismic signals is convenient to be used as a magnitude parameter of pyroclastic flows and rockfalls.

A duration–frequency plot for the seismic signals associated with pyroclastic flows and rockfalls during the 1998 and 2004 lava eruptions at Volcán de Colima was constructed as shown in [Fig. 13.16](#). [Gutenberg and Richter \(1956\)](#) showed that the relationship between the magnitude of seismic events and their frequency characterizes the physical conditions of material fracturing. Pyroclastic flow and rockfall earthquakes are the products of the destruction of lava bodies. A log–log plot of the duration intervals

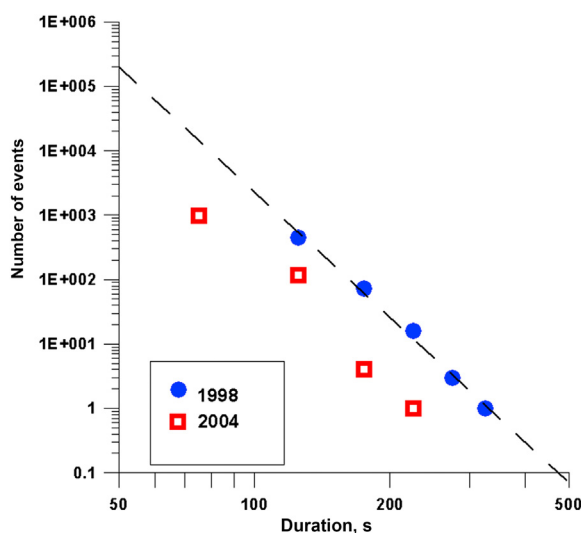


**FIGURE 13.16** Relationship between the number of rockfalls  $N$ , and the duration of their signals,  $D$ , during the 1998 extrusion (circles) and the 2004 extrusion (squares) at Volcán de Colima. The least-square regression line  $\log N = 15.7 - 6.2 \log D$  (s) for the 1998 events is shown. From Zobin, V.M., Varley, N.R., González, M., Orozco, J., Reyes-Dávila, G., Navarro, C., et al., 2008. Monitoring the 2004 andesitic block-lava extrusion at Volcán de Colima, México from seismic activity and  $\text{SO}_2$  emission. *Journal of Volcanology and Geothermal Research* 177, 367–377.

versus the number of events of this duration evidences a robust linear dependence between the parameters. This relationship indicates a power-law distribution of the seismic signals related to pyroclastic flow and rockfall earthquakes occurring due to the partial collapse of the lava dome. This relationship is typical for fractal sets (Scholz, 1990) and may indicate the self-similarity in the dimensions of the pieces that were broken off during the gravitational destruction of the lava body.

### 13.5.2 Relationship Between the Magnitude of Earthquakes Associated With Pyroclastic Flow and Rockfall and the Volume of Pyroclastic Flows Emplaced From Partial Collapse of the Lava Dome at Volcán De Colima, Mexico

The results presented in the previous section suggest that the duration of the short-period records of pyroclastic flow and rockfall earthquakes may provide a way to assess the magnitude of these earthquakes. The duration of the signal was measured beginning from the onset of the signal and till the end of its coda. The following magnitude scale was proposed based on the duration of the rockfall earthquakes observed during the 1998–99 effusive

eruption at Volcán de Colima and recorded by a short-period station situated at a distance of 4 km from the crater:

1. magnitude M1 for events with durations of seismic signal less than 100 seconds (“short” in Fig. 13.14);
2. magnitude M2 for events with durations between 100 and 250 seconds (“inter” in Fig. 13.14); and
3. magnitude M3 for events with durations between 250 and 450 seconds (“long” in Fig. 13.14).

The events of magnitude M1 can be attributed mainly to rockfalls; the magnitudes M2 and M3 events may be associated with pyroclastic flows and large rockfalls. Self-similarity in the dimensions of the pieces that were broken off during the gravitational destruction of the lava body allows us to estimate the mean values of the material that was transported by pyroclastic flows of different magnitudes.

The volume of pyroclastic flows that occurred between November 25, 1998, and November 26, 1998 ( $7.2 \times 10^5 \text{ m}^3$ ), and of the total volume of pyroclastic flows that occurred during the eruption in November 1998–January 1999 ( $1.6 \times 10^6 \text{ m}^3$ ) were estimated by Navarro-Ochoa et al. (2002) and Saucedo et al. (2002), respectively. Thus the approximate mean volume of flows corresponding to seismic records of different durations was estimated. For the November 25 and November 26 observations, 3 events with magnitude M3, 123 events with magnitude M2, and 219 events with magnitude M1 were recorded. For the total period of the eruption, 4 events of magnitude M3, 577 events with magnitude 2, and 4939 events with magnitude M1 were recorded. Field observations showed that the magnitude M1 events correspond to rockfalls with a mean volume of  $50 \text{ m}^3$ . A system of two linear equations can be solved to measure the cumulative volume of material associated with events of magnitude M2 and M3:

$$577V_2(\text{m}^3) + 4V_3(\text{m}^3) = 1.6 \times 10^6(\text{m}^3) - 50(\text{m}^3) \times 4939 \quad (13.1)$$

$$123V_2(\text{m}^3) + 3V_3(\text{m}^3) = 7.2 \times 10^5(\text{m}^3) - 50(\text{m}^3) \times 219 \quad (13.2)$$

where  $V_2$  and  $V_3$  are the mean volumes of material corresponding to the pyroclastic flows of magnitudes M2 and M3, respectively. From the above two equations, a mean volume of  $2.0 \times 10^5 \text{ m}^3$  was obtained for events with magnitude M3, and a volume of  $1.0 \times 10^3 \text{ m}^3$  as a mean value for the events with magnitude M2.

This method of estimating volumes is approximate but may be useful for eruption monitoring. The values of  $V_2$  and  $V_3$  calculated at Volcán de Colima are in agreement with the estimates of the volume of pyroclastic flows performed at Unzen volcano, Japan ( $0.6 \times 10^4$  to  $5.7 \times 10^4 \text{ m}^3$ ; Uhira and Takeo, 1994; Yamasato, 1997) and Soufrière Hills volcano, Montserrat (small pyroclastic flows,  $10^3$ – $10^4 \text{ m}^3$ ; medium-sized pyroclastic flows,  $10^4$ – $10^6 \text{ m}^3$ ; Calder et al., 2002).

### 13.5.3 Relation Between the Duration of Broadband Seismic Signals Associated With Pyroclastic Flows Emplaced From Eruption Columns and the Volume of Pyroclastic Flows at Volcan De Colima, Mexico

The volume of pyroclastic flows emplaced from eruption columns of Vulcanian explosions at Volcán de Colima, Mexico during the July 17, 2003 explosion and six explosions of the May 2005 to September 2005 sequence were obtained during field measurements performed by [Muñoz Martínez et al. \(2006\)](#) and [Zobin et al. \(2016\)](#), respectively. The volumes ranged between  $2.3 \times 10^5 \text{ m}^3$  and  $4 \times 10^5 \text{ m}^3$ .

The measurement of the duration of the seismic phase associated with these pyroclastic flows is illustrated in [Fig. 13.17](#). As shown in [Section 13.2.2](#), the fall of volcanic rocks from the eruptive column and generation of pyroclastic flow begins 20–25 seconds after the onset of the seismic signal. In [Fig. 13.17B](#), two phases of the seismic signal are separated. The initial phase of 25 seconds duration, associated with the explosion, is shown within a rectangle; the following part of the signal is considered to be generated by the emplacement of pyroclastic

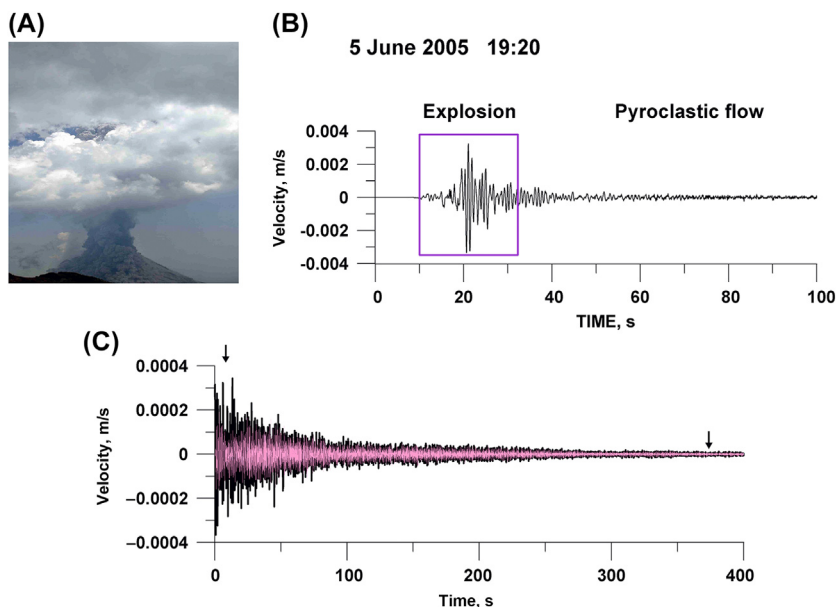


**FIGURE 13.17** Processing of broadband seismic record of the June 5, 2005 explosion at Volcán de Colima (vertical component, 4 km from the crater). (A) Photo of the explosion taken by Guillermo Arias; (B) seismic record of the explosion and following pyroclastic flow (the explosion phase is taken in rectangle); and (C) the filtered seismic signal of pyroclastic flow (see text). Arrows in (C) show the beginning and the end of the seismic signal of pyroclastic flow. *Courtesy Colima Volcano Observatory.*

flow. The selected signal of pyroclastic flow was then high-pass filtered at 1.5 Hz to remove the low-frequency noise produced by the explosion (Fig. 13.17C). The duration of the filtered signal associated with the pyroclastic flow was measured from the beginning of the signal signifying the end of the explosion phase and the end of the signal marked as when its amplitude becomes equal to the background level. The duration of the seismic signals of pyroclastic flows estimated for seven Colima explosions ranged from 270 to 400 seconds.

Fig. 13.18 shows the plot of values of volume of pyroclastic flows emplaced from eruption columns of the Colima explosions versus the durations of respected seismic signals. These data are plotted together with the values of mean volumes of pyroclastic flows emplaced from the partial collapse of the lava body, obtained for the magnitude M2 and M3 seismic signals in Section 13.5.2. The data obtained for pyroclastic flows emplaced from eruption columns (durations of seismic signals from 270 to 400 seconds) are clustered within the interval of the seismic signal durations proposed for the magnitude M3 earthquakes (from 250 to 450 seconds). The values of volume of these pyroclastic flows only slightly exceed the mean volume of pyroclastic flows emplaced from the partial collapse of the lava body corresponding to the

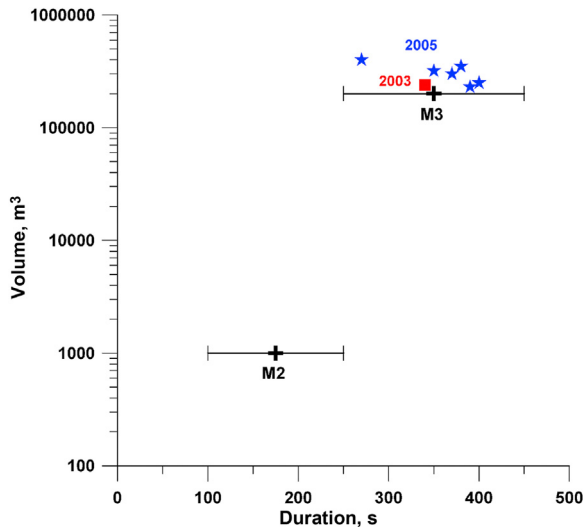


**FIGURE 13.18** Durations of seismic signals corresponding to the movement of pyroclastic flows emplaced from eruption columns at Volcán de Colima versus the volumes of these flows. The volume measurements for the 2005 events, performed by Victor Vargas and Nick Varley (Zobin et al., 2016a), are shown with blue stars. Data obtained by Muñoz Martínez et al. (2006) for the explosion of July 17, 2003 are shown with red rectangle (gray in print versions). Crosses labeled M2, and M3 correspond respectively to the mean volumes of pyroclastic flows emplaced from the partial collapse of the lava body for the seismic signal duration intervals between 100 and 250 s (magnitude M2), and between 250 and 450 s (magnitude M3; see Section 13.5.2). The signal duration intervals for the events of magnitudes M2 and M3 are shown. *Courtesy Colima Volcano Observatory.*

magnitude M3 events. This makes comparable the volumes of both types of pyroclastic flows if their seismic signal durations are closed.

### 13.5.4 Relationship Between the Amplitude of Long-Period Seismic Records of Pyroclastic Flow and Rockfall Earthquakes and the Volume of Pyroclastic Flows at Unzen Volcano, Japan

Uhira and Takeo (1994) showed that the maximum amplitudes of long-period seismic records,  $A_{\max}$ , produced by pyroclastic flows emplaced by the partial collapse of the lava dome at Unzen volcano during the 1991–92 eruption were proportional to the volume of the blocks fallen from the dome,  $V$  (Fig. 13.19). The seismograms were recorded at a distance of about 4 km from the lava dome.

The least-square regression between these two parameters was calculated based on Fig. 13.22 from Uhira and Takeo (1994): (13.3)

$$V(10^5 \text{ m}^3) = 0.22 A_{\max}(\text{micron}) + 0.06 \quad (13.3)$$

The correlation coefficient is equal to 0.93; therefore, the maximum amplitude of the long-period signal may also be used for the quantification of pyroclastic flows.

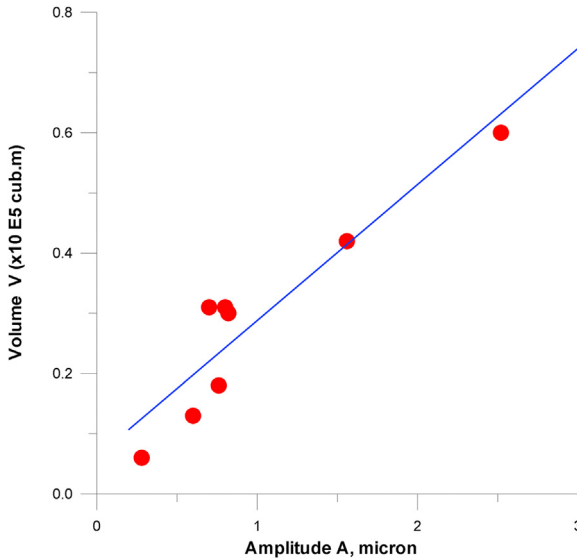


**FIGURE 13.19** Relationship between the volume of pyroclastic flows and the maximum amplitude of long-period records (vertical component) of the earthquakes produced by pyroclastic flows at Unzen volcano, Japan, in 1991–92. The seismic station was situated at a distance of about 4 km from the crater. *Modified from Uhira, K., Yamasato, H., Takeo, M., 1994. Source mechanism of seismic waves excited by pyroclastic flows observed at Unzen volcano, Japan. Journal of Geophysical Research 99, 17757–17773.*

### 13.6 TRACKING OF PYROCLASTIC FLOWS TRAJECTORY USING THE AMPLITUDE SIGNALS OF EARTHQUAKES

The amplitude of the signals recorded by a high dynamic range seismograph network may be used for tracking the trajectory of pyroclastic flows and rockfalls (Yamasato, 1997; Jolly et al., 2002). Fig. 13.20 shows typical seismograms and spectrograms for an earthquake recorded

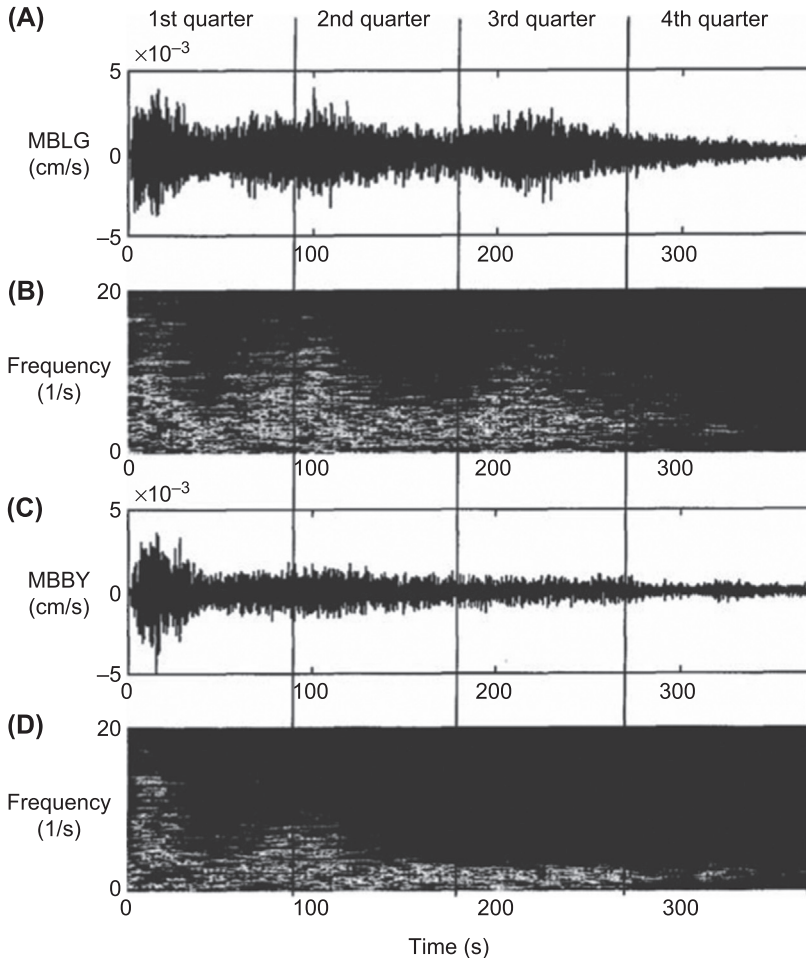


**FIGURE 13.20** Seismograms (A, C) and spectrograms (B, D) of pyroclastic flow recorded on August 12, 1999, at Soufrière Hills volcano, Montserrat at a distance of about 2 km NE from the dome (station MBLG), and about 2 km SW from the dome (station MBBY). From Jolly, A.D., Thompson, G., Norton, G.E., 2002. Locating pyroclastic flows on Soufrière Hills volcano, Montserrat, West Indies, using amplitude signals from high dynamic range instruments. *Journal of Volcanology and Geothermal Research* 118, 299–317.

at Soufrière Hills volcano, Montserrat, and associated with the emplacement of a pyroclastic flow.

The signal observed at one or more seismic stations reflects the onset of the flow at a single location, followed by multiple moving sources spread over a large area. The amplitudes of these sources can be exploited to determine an average location for the event. Jolly et al. (2002) used the spectral amplitudes of the signal at a frequency of 7–9 Hz to track pyroclastic flow propagation. The seismic record was divided into four segments (Fig. 13.20), and locations were calculated for each segment. Fig. 13.21 shows locations of the trajectory of propagation of a pyroclastic flow on August 12, 1999, at Soufrière Hills volcano. It can be seen that the pyroclastic flow propagated from the dome to NE with a run out of about 1.4 km.

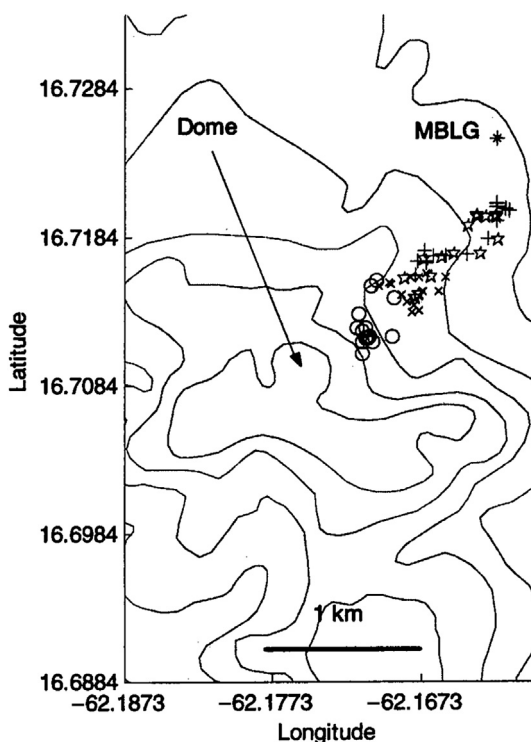


**FIGURE 13.21** Amplitude-based locations by a network of five seismic stations for the pyroclastic flow recorded on August 12, 1999, at Soufrière Hills volcano, Montserrat. The event is divided into four time intervals to show propagation through time (see Fig. 13.20). The first segment is marked by circles, the second by crosses, the third by plus signs, and the fourth by stars. The amplitude-based locations suggest that the event was initiated on the southern part of the dome and subsequently propagated along the Tar River Valley. From Jolly, A.D., Thompson, G., Norton, G.E., 2002. Locating pyroclastic flows on Soufrière Hills volcano, Montserrat, West Indies, using amplitude signals from high dynamic range instruments. *Journal of Volcanology and Geothermal Research* 118, 299–317.



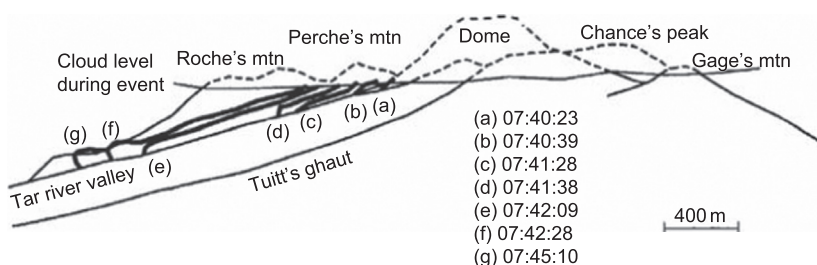


**FIGURE 13.22** Schematic diagram extracted from video footage recorded on August 12, 1999, at Soufrière Hills volcano, Montserrat, showing pyroclastic flow trajectory through time. The view of the video camera is to the north from the dome. From Jolly, A.D., Thompson, G., Norton, G.E., 2002. Locating pyroclastic flows on Soufrière Hills volcano, Montserrat, West Indies, using amplitude signals from high dynamic range instruments. *Journal of Volcanology and Geothermal Research* 118, 299–317.

The comparison with the video camera record of the pyroclastic flow (Fig. 13.22) strongly supports the seismic amplitude-based reconstruction of the movement of pyroclastic flow.

## 13.7 SEISMIC SIGNALS ASSOCIATED WITH LAHARS: WAVEFORMS AND SPECTRA

A lahar, or volcanic mudflow, is a turbulent-muddy debris flow or a hyperconcentrated flood flow on volcanic slopes irrespective of temperature, in which heavily ejected fine ash or pyroclastic materials are the main constituents (see Section 11.1.2). Lahars are characterized by the fronts enriched in blocks (FEBs) followed by the main flow body, the dilute surges (DS) from flow discharge fluctuations, and dilute tails. Lahars, having only one FEB, are classified as single-pulse events and, having a sequence of FEB, as multi-pulse events (Vázquez et al., 2016). A lahar starts as a debris flow for the first few kilometers but then evolves into a hyperconcentrated flood flow (Takahashi, 2007).

### 13.7.1 Volcán De Colima, Mexico

#### 13.7.1.1 Lahars at Volcán de Colima

Lahars at Volcán de Colima occur during the rainy seasons, between June and October. They are formed when the rainy water is saturated with the blocks of pyroclasts and ash collected during the eruptive activity at Volcán de Colima. The lahars at Volcán de Colima usually went along two ravines, Montegrando and La Lumbre (Fig. 13.23). These two ravines are differing in their morphology. La Lumbre is characterized by a more straight drainage system compared with the strongly curved Montegrando ravine. Both ravines are characterized by the vertical boards with a width changing from about 10 m (gradient of about 32°) at a distance of about 1.6 km from the crater of Colima volcano to about

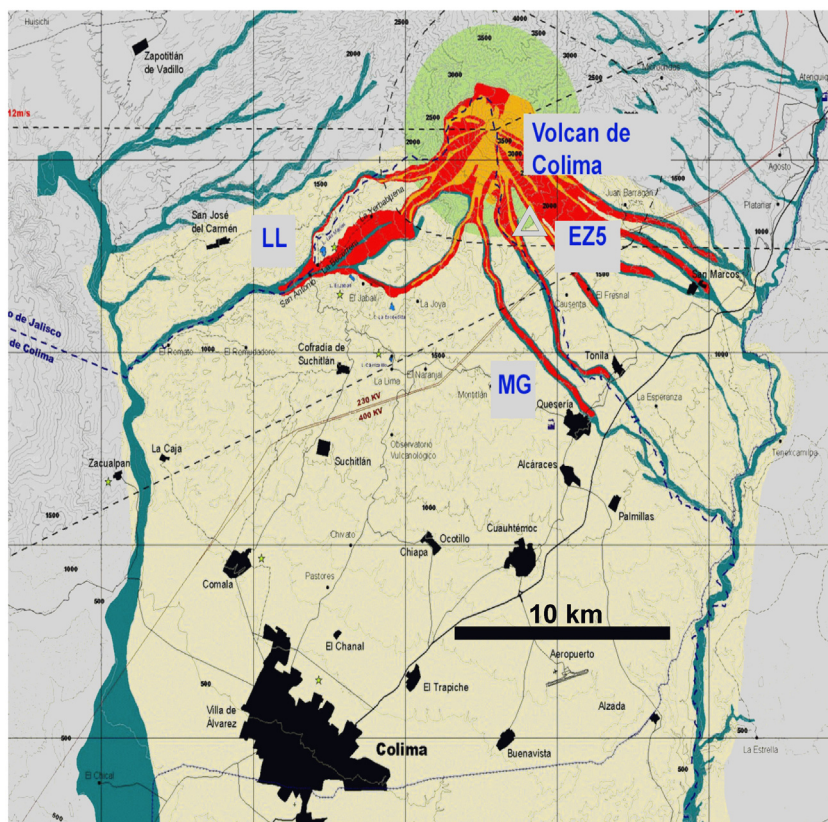


**FIGURE 13.23** A portion of the map of volcanic hazards of Volcán de Colima showing the ravines of Montegrande (MG) and La Lumbre (LL) and the position of broadband seismic station EZ5 shown as a triangle. Green (dark gray in print versions) zones show the ravines where lahars propagate. The map of volcanic hazards of Volcán de Colima, prepared by C. Navarro and A. Cortés, is taken from the Website of Centro Universitario de Estudios Vulcanológicos (<http://portal.ucol.mx/cueiv/mapapeligros.htm>).

20 m (gradient of about  $19-7^\circ$ ) at a distance of about 10 km. Then their channels widen from 40 to 100 m (Davila et al., 2007; Abel Cortes, personal communication, 2008). The peak flow velocity of some lahars was estimated from the video records as about 10 m/s. They reach about 8–12 km from the volcanic cone. The rain gauge, installed in 2006 near the Montegrande ravine at the height of 2450 m a.s.l., showed that the triggering of lahars occurred if the rain intensity had reached 1–2 mm/min.

### 13.7.1.2 System of Monitoring

The seismic signals associated with lahars, occurring during 2005–13, were recorded by broadband seismic station EZ5 installed at a distance of 4 km

from the crater near the Montegrande ravine (Fig. 13.23). During monitoring of lahars in 2012 and 2013, a video camera Axis M1113-E, which took images every 2 seconds with a resolution of  $704 \times 480$  pixels, was installed at a distance of 5.8 km to the south of the crater and fixed 3 m from the ground and 13 m from the channel of the Montegrande ravine (Vázquez et al., 2016).

### 13.7.1.3 Waveforms of the Seismic Signals of Lahars

The duration of seismic records produced by lahars is substantial, up to a few hours. Seismic signals reflect the main stages in the lahar movement. Fig. 13.24 shows broadband seismic records associated with the propagation of two lahars developed on the slope of Volcán de Colima on June 11, 2013, and June 25, 2012, and June 25,

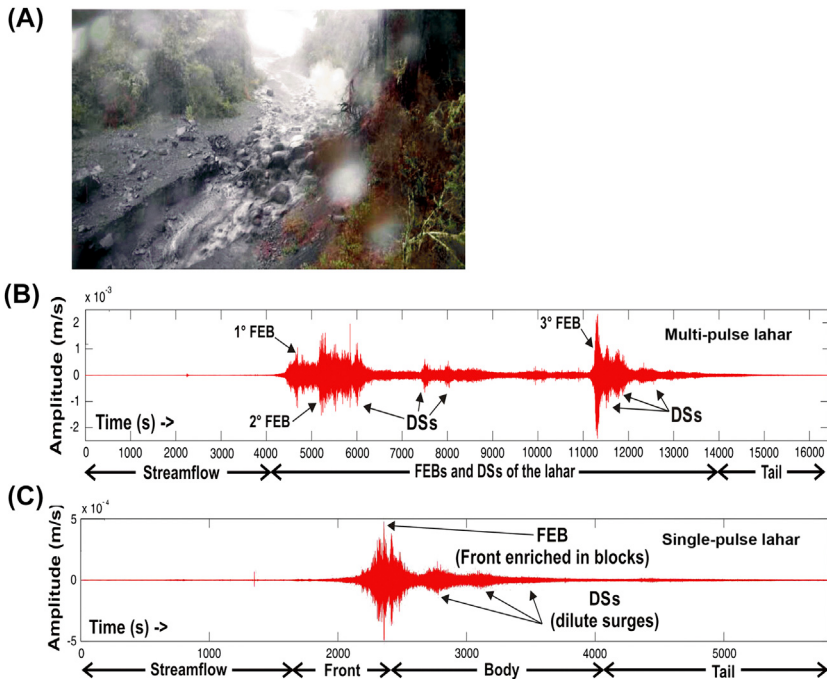


**FIGURE 13.24** Broadband seismic records (velocity, vertical components, station EZ5) of the multipulse June 11, 2013 (B) and the single-pulse June 25, 2012 (C) lahars at Volcán de Colima. In (A) a video image of the block-rich front of the June 11, 2013 lahar recorded along the Montegrande ravine by the lahar monitoring station at a distance of 5.8 km to the south of the crater is shown. DS, Dilute surges; FEB, block-rich front. *Courtesy Centro de Geociencias, UNAM. Figures B and C are modified from Vázquez, R., Suriñach E., Capra, L., Arámbula-Mendoza, R., Reyes-Dávila, G., 2016. Seismic characterisation of lahars at Volcán de Colima, Mexico. Bulletin of Volcanology 78 (8), 1–14 and Zobin, V.M., Plascencia, I., Reyes, G., Navarro, C., 2009a. The characteristics of seismic signals produced by lahars and pyroclastic flows: Volcán de Colima, México. Journal of Volcanology and Geothermal Research 179, 157–167.*

2012, and passing along the Montegrando ravine (see Fig. 13.23). Simultaneous seismic and video monitoring of these events allowed Vázquez et al. (2016) to discriminate the different elements of lahars within the seismic signals such as the FEB, DS, and dilute tails (Fig. 13.24B and C).

The June 11, 2013 lahar had more than one FEB, and it was classified as a multipulse lahar. The first FEB of the lahar (marked as 1° FEB on the seismogram in Fig. 13.23B), recorded at a distance of 5.8 km to the south of the crater of Volcán de Colima, is shown in Fig. 13.24A. A sequence of seismic signals reflects the complex nature of this event and is associated with the arrival of the FEBs and DSs. Three FEBs and six DSs, identified according to the recorded images, are good corresponding to the variations in the amplitudes and shape of seismic signals. The propagation of lahar lasted approximately 3 hours. The calculated flow discharge for the first FEB was  $26 \text{ m}^3/\text{s}$ , and  $60 \text{ m}^3/\text{s}$  for the second FEB.

The single-pulse lahar, occurring on June 25, 2012, was characterized by a single FEB followed by three DSs, visible as flow discharge fluctuations within the recorded images and identified by the shape and amplitudes of the seismic signals (Fig. 13.24C). The FEB was associated with a peak discharge of  $20 \text{ m}^3/\text{s}$ . The event lasted approximately 1 hour. The seismic waveforms associated with this type of lahar represent the emergent, cigar-shaped waxing, and waning seismic signals that gradually increase in time during the passing of the FEB of lahar reaching their maximum amplitude and then gradually decreasing. The secondary signals, associated with the DS, are similar in shape but have lower amplitudes compared with the FEB phase.

#### 13.7.1.4 *Influence of Morphology of the Ravines on the Seismic Waveforms*

The influence of the morphology of the ravines on the seismic waveforms associated with the single-pulse lahars is illustrated in Fig. 13.25 showing the 1-hour seismic records produced by lahars originated at Volcán de Colima during the 2005–06 activity and passing along the ravines La Lumbre (Fig. 13.25A) and Montegrando (Fig. 13.25B) situated on the flanks of the volcano (see Fig. 13.23). Both groups of records have similar waveforms. A detailed analysis of the initial records of lahars is practically impossible due to the large noise that hides the small amplitude initial impulses produced by the initial stage of a lahar. Nevertheless, it is possible to separate some groups of impulses. For the case of the 2006\_07150414 La Lumbre lahar (Fig. 13.26A), the first low-amplitude impulse has a duration of about 8 minutes. The onset of the impulse 2 is sharp; it has a larger amplitude than the impulse 1. Impulse 2 gradually decreased over about 40–50 minutes. The record of the 2005\_0506300500 Montegrando lahar (Fig. 13.26B) consists of at least three impulses with increasing amplitude.

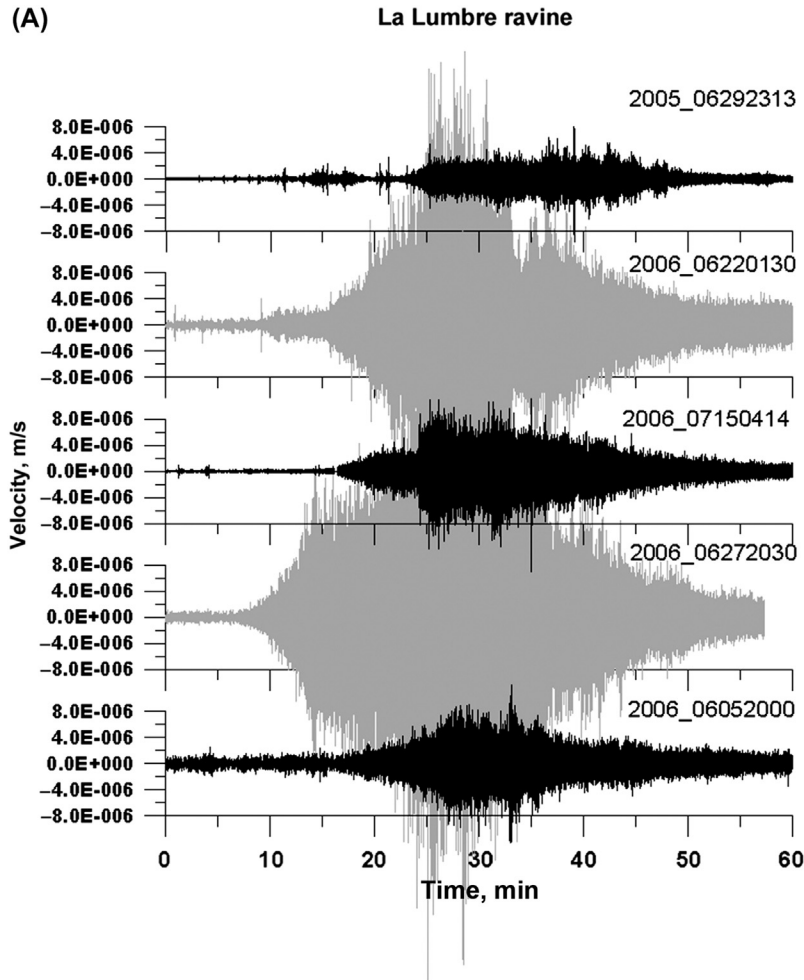


**FIGURE 13.25** Broadband seismic velocity records (station EZ5, vertical component) of lahars propagating along the La Lumbre (A) and Montegrande (B) ravines on the flanks of Volcán de Colima. The index near seismic record represents the date (yyyy\_mmddhhmm) of the events.

The final long impulse 3 is similar to the final long impulse 2 of the La Lumbre event. The presence of sharp, large amplitude, long-duration final impulses may indicate a change in the nature of lahars during their propagation or a change in their path conditions.

### 13.7.1.5 Spectral Content

Fig. 13.27 shows the spectral characteristics of broadband seismic signals of lahars illustrated in Fig. 13.24 and Fig. 13.25. The spectral content of

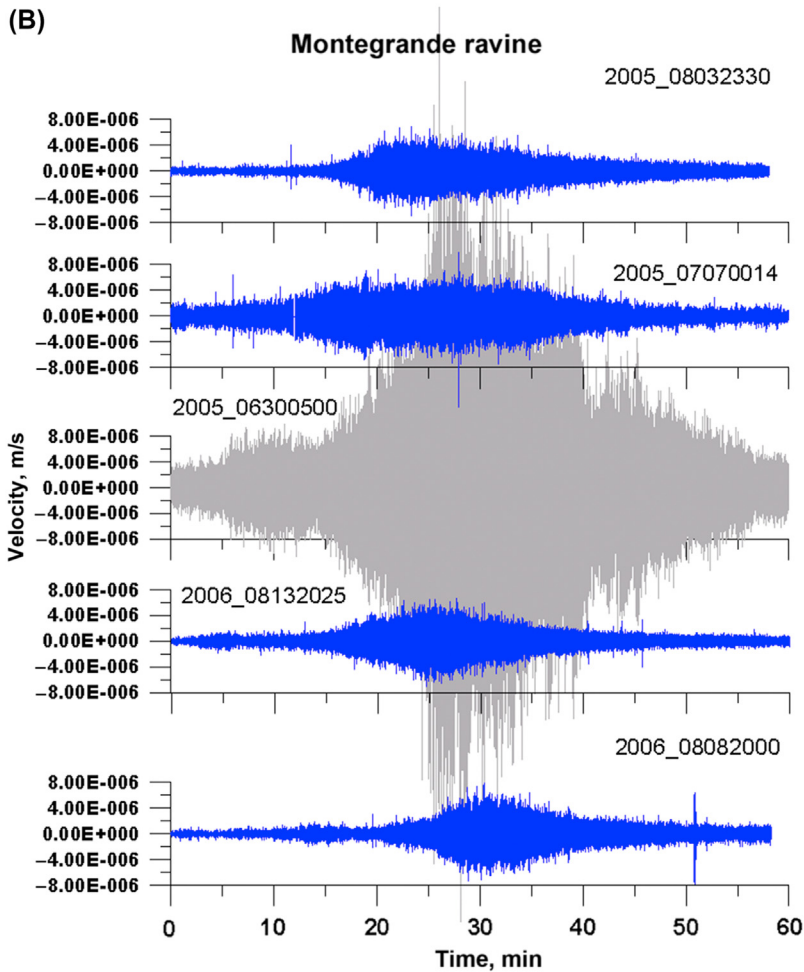


FIGURE 13.25 (Continued)

the seismic pulses corresponding to the FEBs and DSs of lahars is different (Fig. 13.27A and B). The frequencies of the seismic signals, corresponding to the FEB, ranged between 5 and 15 Hz. The frequencies of the seismic signals corresponding to the DS ranged between 10 and 30 Hz. No special features due to different relief of the ravines are seen on Fourier spectra calculated for broadband seismic records of two groups of lahars passing along two ravines, La Lumbre and Montegrando (Fig. 13.27C). They are similar with the peak frequencies between 6.3 and 7.6 Hz corresponding to the characteristic frequencies of the seismic signals associated with the FEB.



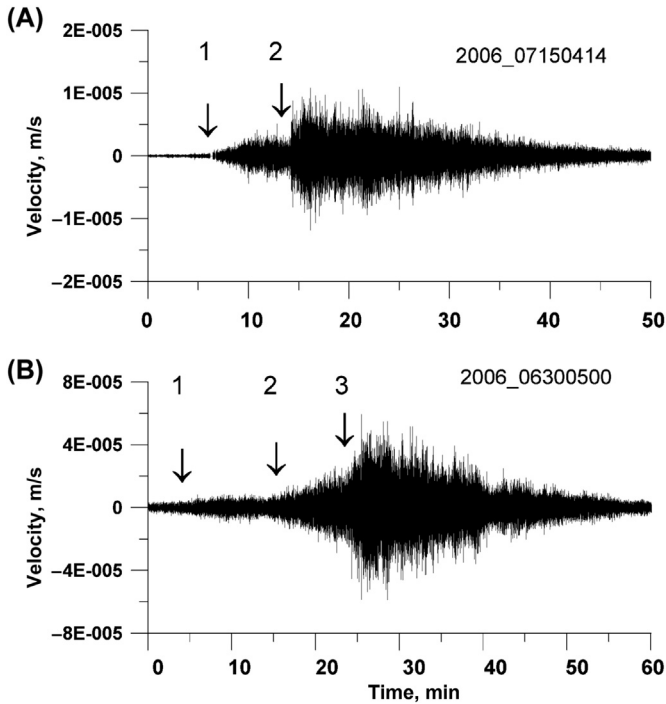


**FIGURE 13.26** Structure of broadband seismic records (vertical component) of the 2006\_07150414 La Lumbre lahar (A) and the 2005\_0506300500 Montegrande lahar (B) shown in Fig. 13.25. Arrows with number show the arrivals of the seismic impulses (see text). *Courtesy Colima Volcano Observatory.*

### 13.7.2 Tungurahua Volcano, Ecuador

Tungurahua is an andesitic stratovolcano in the Ecuadorian Andes. This volcano has continued eruptive activity since 1999, in which major eruptions were characterized by Vulcanian and Strombolian eruptive styles accompanied by lahars. Six broadband seismic stations situated at distances of 2–6 km from the crater with a wide azimuthal coverage were in operation (Fig. 13.28A). Kumagai et al. (2015) used the amplitude source location method to estimate the positions of the source of lahars considering the source amplitude to be the maximum value during the seismic record of lahar. Seven sources of lahars were located on the northern and northwestern slopes of the volcano. The signals were characterized by similar long spindle shapes with durations from 30 to 50 minutes.

Fig. 13.28B demonstrates broadband seismic signals associated with the lahar that occurred on February 17, 2011. They indicate the strong attenuation of the lahar records with distance. Only the two nearest seismic stations, BBIL and BMAS, installed at a distance of about 2 km from the located source, represent good-quality signals.

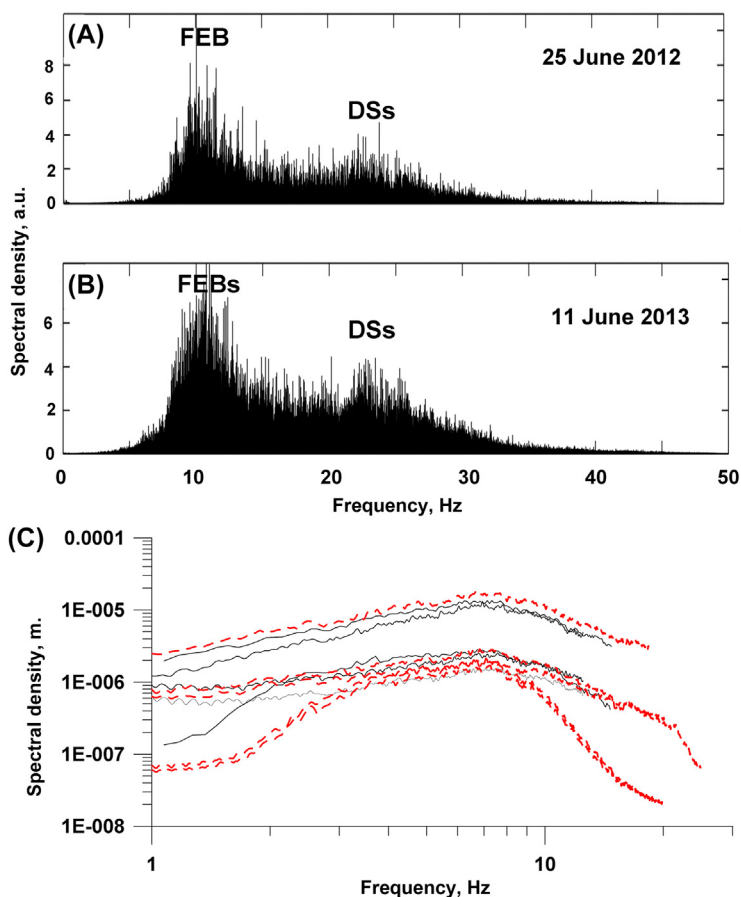


**FIGURE 13.27** Spectra of broadband seismic records of lahars shown in Figs. 13.24 and 13.25. (A) The June 25, 2012, single-pulse event, (B) the June 11, 2013 multipulse event, and (C) the lahars propagating along the La Lumbre (*thin black lines*) and Montegrande (*dashed red lines* (gray in print versions)) ravines. Modified from Vázquez, R., Suriñach E., Capra, L., Arámbula-Mendoza, R., Reyes-Dávila, G., 2016. Seismic characterisation of lahars at Volcán de Colima, Mexico. *Bulletin of Volcanology* 78 (8), 1–14 and Zobin, V.M., Plascencia, I., Reyes, G., Navarro, C., 2009a. The characteristics of seismic signals produced by lahars and pyroclastic flows: Volcán de Colima, México. *Journal of Volcanology and Geothermal Research* 179, 157–167.

### 13.7.3 Mt. Merapi Volcano, Indonesia

Mt. Merapi is an andesitic stratovolcano located in Central Java. Since 1992, Merapi experienced quasi-continuous extrusions of lava at its summit, which forms a dome in a horseshoe-shaped crater. More than 50 lahars were generated around Mt. Merapi during the rainy season following the *nuées ardentes* of November 22, 1994 (Lavigne et al., 2000).



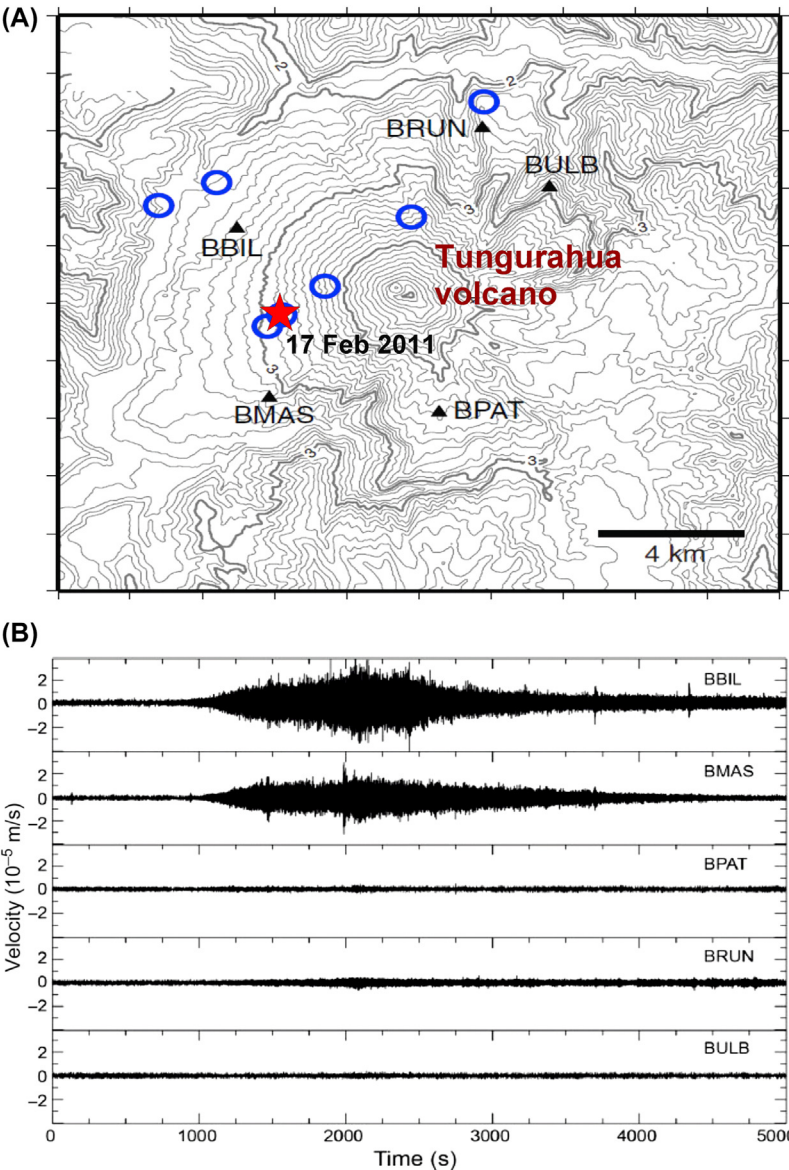


**FIGURE 13.28** Network of six broadband seismic stations around Tungurahua volcano (A) and the seismic signals (velocity, vertical components) produced by the February 17, 2011 lahar (B). In (A), triangles show the positions of seismic stations, circles show the locations of the sources of lahars, and a star indicates the location of the source of the February 17, 2011 event. Modified from Kumagai, H., Mothes, P., Ruiz, M., Maeda, Y., 2015. An approach to source characterization of tremor signals associated with eruptions and lahars. *Earth Planet Space* 67 (178), 1–12.

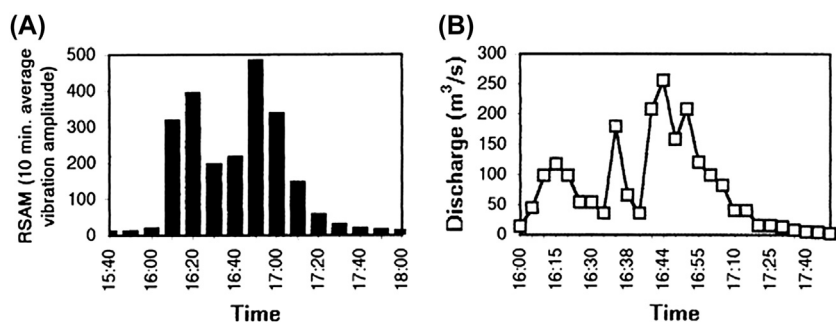


**FIGURE 13.29** RSAM data (A) and observed discharge (B) of the February 19, 1995 lahar at Mt. Merapi volcano. Observations were performed on the elevation of 805 m at a distance of 7.8 km from the crater. *Modified from Lavigne, F., Thouret, J.C., Voight, B., Young, K., LaHasen, R., Marso, J., et al., 2000. Instrumental lahar monitoring at Merapi volcano, Central Java, Indonesia. Journal of Volcanology and Geothermal Research 100, 457–478.*

Lahar dynamics were monitored using RSAM. The best-frequency band for monitoring the Merapi lahars ranged from 8.75 to 9.25 Hz. Lahar discharge at Merapi was well correlated with the RSAM amplitudes. RSAM data representing 10-minute average amplitudes were compared and correlated to the instantaneous discharge variations of the lahars. The highest RSAM value (484 units) was recorded in connection with the February 19, 1995 lahar (Fig. 13.29).

### 13.8 COMPARISON OF THE SEISMIC CHARACTERISTICS OF PYROCLASTIC FLOWS AND LAHARS

Fig. 13.30 represents the seismic characteristics of pyroclastic flow and lahars, recorded at a distance of 4 km from the crater of Volcán de Colima, in terms of the relationships “duration of seismic signal versus spectral peak frequency.” It is clearly seen that the groups of characteristics for each type of volcanic flow are separated. The second-to-minute durations of the seismic signals of pyroclastic flow significantly differ from the 10-minute-to-hour durations of the seismic signals of lahars. The peak frequencies of Fourier spectra of pyroclastic flows (2–4 Hz) are significantly lower than the same of lahars (6–8 Hz).

This observation allows a simple recognition of these events in real time. A spindle-shaped signal, completely recorded on a seismogram during the first 3–5 minutes, would indicate a pyroclastic flow but if a signal gradually increases in amplitude during the same time interval; it may refer to the beginning of lahar movement. It may be used for early diagnostics of the type of volcanic flow in real time.

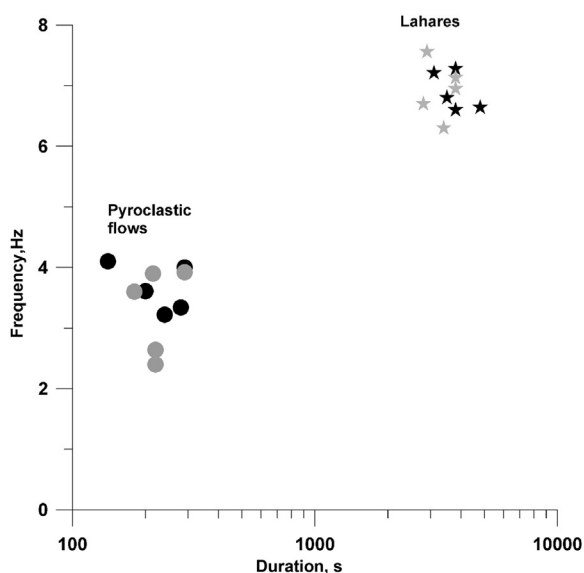


**FIGURE 13.30** Durations versus spectral peak frequencies of broadband seismic signals associated with lahars and pyroclastic flows occurring at Volcán de Colima and shown in Figs. 13.3, 13.7, and 13.25. All seismic signals were recorded at a distance of 4 km from the crater. Characteristics of the seismic signals associated with pyroclastic flows emplaced from eruption columns are shown with *black circles* and those of pyroclastic flows emplaced from partial collapse of the lava dome are shown with *gray circles*. Characteristics of the seismic signals associated with lahars, propagating along the La Lumbre ravine, are shown with *gray stars*, whereas those propagating along the Montegrando ravine, are shown with *black stars*. *Courtesy Colima Volcano Observatory.*

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## RELEVANT WEBSITE

BGVN. Bulletin of Global Volcanic Network. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.

## Chapter 14

# Seismic Signals Associated with Volcanic Explosions

The most spectacular eruptions consist of violent explosions that blast great clouds of gas-laden debris into the atmosphere. The three most disastrous volcanic eruptions during the past two centuries were explosive with a number of victims from 30,000 to 250,000 fatalities ([https://en.wikipedia.org/wiki/List\\_of\\_volcanic\\_eruptions\\_by\\_death\\_toll](https://en.wikipedia.org/wiki/List_of_volcanic_eruptions_by_death_toll)). As was noted in Chapter 1, the main types of volcanic magmatic explosions are Strombolian (VEI 1–2), Vulcanian (VEI 2–4), Plinian (VEI 4–8), and Ultra-Plinian (VEI 5–8). Strombolian explosions are short (seconds to minutes), discrete, and produce eruptive columns or lava fountains; they are the characteristics for basaltic volcanoes and occur at quasiregular intervals producing 3–5 explosions/h with the eruption columns of tens or hundreds of meters in height (Vergnolle and Mangan, 2000). Vulcanian explosions are also short (seconds to minutes) discrete explosions with a small amount (less than 1 km<sup>3</sup>) of erupted products but they are characterized by high eruption columns, which can reach 10–20 km altitude, and occupy the intermediate position in the explosion hierarchy between low-rise Strombolian and high-rise Plinian explosions whose eruption columns may reach up to 58 km altitude (Francis, 1998; Morrissey and Mastin, 2000; Schmincke, 2004; Yuen et al., 2022). Vulcanian and Plinian explosions are the characteristics of andesitic and dacitic volcanoes. They may be violent, sometimes destroying a part of the volcano edifice and producing highly fragmented ash (Francis, 1998; Morrissey and Mastin, 2000; BGVN, 2022).

A phreatomagmatic explosive eruption is driven primarily by the volumetric expansion of external water after it has been rapidly heated by contact with magma. The most commonly recognized styles of explosive events occurring during phreatomagmatic eruptions are the explosive jets of ash accompanied by billowing clouds of steam (so-called Surtseyan explosion) or Vulcanian-type explosions (Morrissey and Mastin, 2000).

The explosive events, occurring during magmatic, phreatic, or phreatomagmatic eruptions, generate seismic signals. Explosion earthquakes may occur as a single event or as an earthquake sequence. Tokarev and Firstov (1967) studied the seismic activity of the Karymsky volcano, Kamchatka in 1965, when small

explosions occurred, accompanied by a slow emission of dacitic lava. During 3 weeks of observation, a total of 727 explosion earthquakes were recorded. [Rowe et al. \(2000\)](#) described the swarms of seismic signals associated with the passive degassing and Strombolian explosions that originated within a summit lava lake at Mount Erebus, Antarctica. They typically occurred several times daily and reached up to 900 events/day. A short-period 10-minute seismogram recorded at Aso volcano, Japan during the 2015 eruption episode of Strombolian activity ([Fig. 14.1A](#)) illustrates the numerous seismic signals of explosive degassing and small Strombolian explosions. A short-period 24-hour seismogram recorded at Volcán de Colima, Mexico during the May 2015 episode of the lava dome growth demonstrates the seismic signals associated with a sequence of Vulcanian explosive events ([Fig. 14.1B](#)).

## 14.1 WAVEFORMS AND SPECTRA

The seismic records of explosion earthquakes are characterized by a great variety. At the same time, they may form families of similar waveforms ([Johnson et al., 1998](#); [Rowe et al., 2000](#)) depending on the type and style of explosive activity. The seismic records of explosive events may be complicated by the following signals of pyroclastic flows (see Chapter 13) or acoustic signals (see Chapter 18).

### 14.1.1 Strombolian Explosions

Strombolian explosions are the characteristic of silica-poor magmas, often of basaltic to basaltic-andesite composition ([Taddeucci et al., 2015](#)). The seismic signals, recorded at volcanoes Stromboli and Aso and the characteristics for this type of volcanic explosions, are presented in this section.

#### 14.1.1.1 Stromboli Volcano

The volcanic explosions recorded at this volcano gave a name to this type of explosion. The seismic signals associated with Strombolian explosions represent two-phase records. A typical short-period seismogram of Strombolian explosion (vertical component of ground velocity), recorded during the 1992 activity at Stromboli volcano at a distance of 150 m from the crater ([Chouet et al., 1997](#)), showed an initial low-frequency (LF) phase with dominant frequency near 2 Hz, and the main phase of higher frequency up to 10 Hz starting roughly 1 second after the onset of the event ([Fig. 14.2](#)). The onset of the higher frequency signal was noted to coincide with the visually observed onset of the eruptive jet, while no visible surface activity could be linked with the precursory LF onset. The acoustic signal, shown in [Fig. 14.2](#), was characterized by an impulsive arrival at about 2 seconds after the initial onset of the seismic signal. [Martini et al. \(2007\)](#) described the similar seismic signals associated with the 2007 Strombolian explosions at Stromboli volcano.



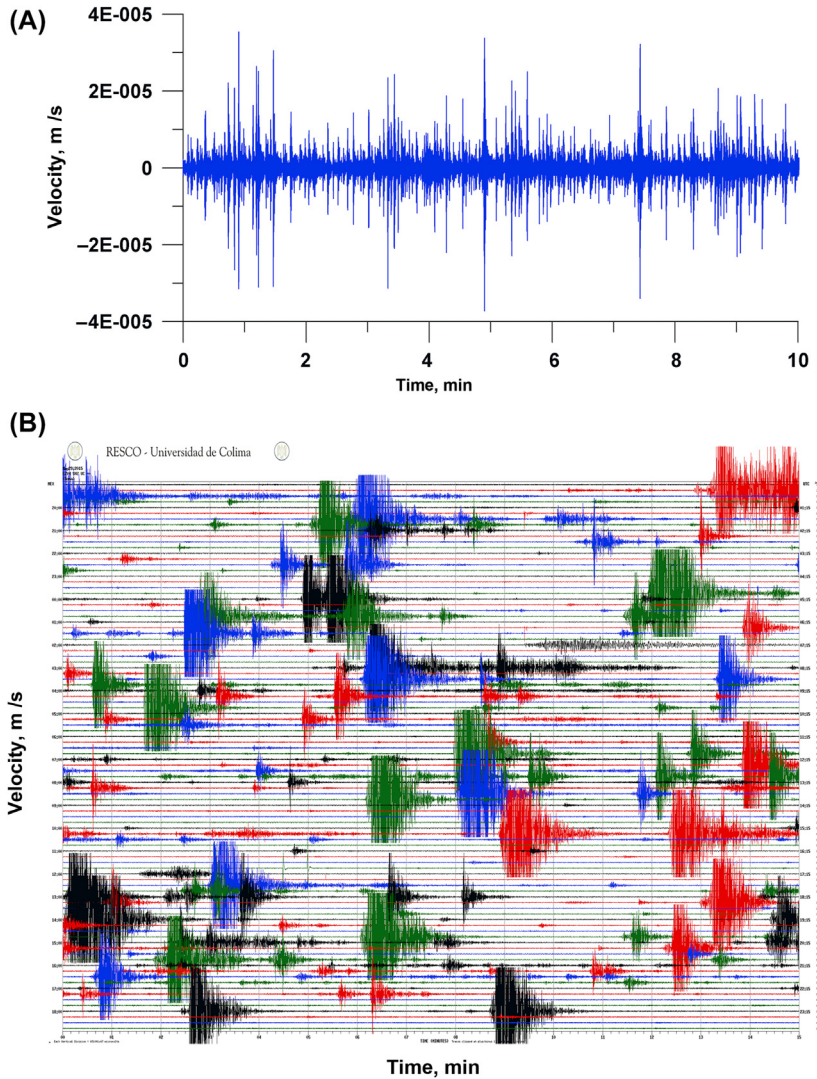


**FIGURE 14.1** (A) Short-period 10-min seismogram (velocity, vertical component) recorded during the 2015 eruption episode of Strombolian activity at Aso volcano at a distance of 150 m from the crater and illustrating the occurrence of the seismic signals generated by degassing and small explosions. Seismic station was situated at a distance of 0.05 km from the rim of the crater. (B) Short-period 24-h seismogram (velocity, vertical component) recorded at Volcán de Colima, Mexico at a distance of 1.7 km from the crater during the May 2015 episode of the lava dome growth demonstrates the seismic signals associated with a sequence of Vulcanian explosive events. Each horizontal line corresponds to a 15-min record. *Courtesy: Aso Volcano Museum (A) and Colima Volcano Observatory (B).*



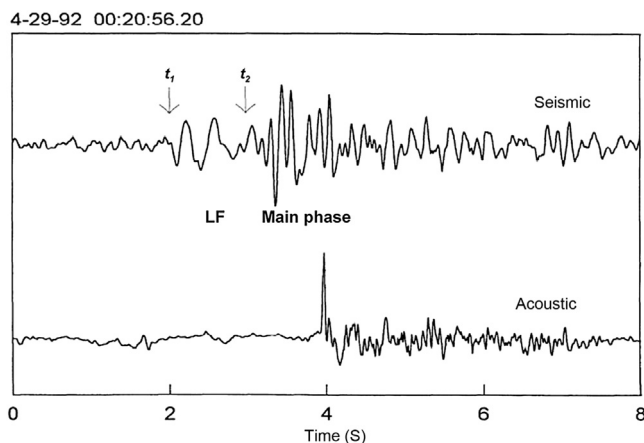


**FIGURE 14.2** Seismic (velocity, vertical component) and acoustic signals associated with Strombolian explosion at Stromboli volcano, Italy (April 29, 1992, 00:20). Short-period seismic station SS1 together with the acoustic instrument was located at a distance of 0.15 km from the crater. LF and main phase indicate the initial low-frequency phase and higher-frequency main phase of the seismic signal, respectively. Individual phase arrivals are identified by the arrows labeled  $t_1$  and  $t_2$ . After Chouet, B., Saccorotti, G., Martini, M., Dowson, P., De Luca, G., Milana G., Scarpa, R., 1997. Source and path effects in the wave fields of tremor and explosions at Stromboli volcano, Italy. *Journal of Geophysical Research* 102, 15129–15150.

They consider that the two components of the seismic radiation, noted by Chouet et al. (1997), are linked to two different source processes during an explosion (see also Section 3.2.2.4, Image 3.9 from Chapter 3).

#### 14.1.1.2 Aso Volcano

Fig. 14.3 shows the short-period seismic signal and its Fourier spectrum associated with the Strombolian explosion, recorded at Aso volcano, Japan at a distance of 50 m from the rim of the crater in December 2014, together with the photo of this explosion. The seismic signal is also represented by an initial LF signal with a duration of 0.4 seconds and the main phase of higher frequency. The Fourier spectrum of the LF phase has a dominant frequency of 7.7 Hz, and the main phase demonstrates a spectral peak of the signal at 12 Hz (Fig. 14.3C).

### 14.1.2 Vulcanian Explosions

Fig. 14.4 shows four typical broadband seismic waveforms (three components) associated with Vulcanian explosions occurring at Volcán de Colima, Mexico, together with the video images of corresponding explosions. As for Strombolian explosions, all these waveforms are characterized by the existence of a large-amplitude, high-frequency (HF) signal of the main phase. The main phase is preceded by an LF signal (Type 1), HF signal (Type 2),

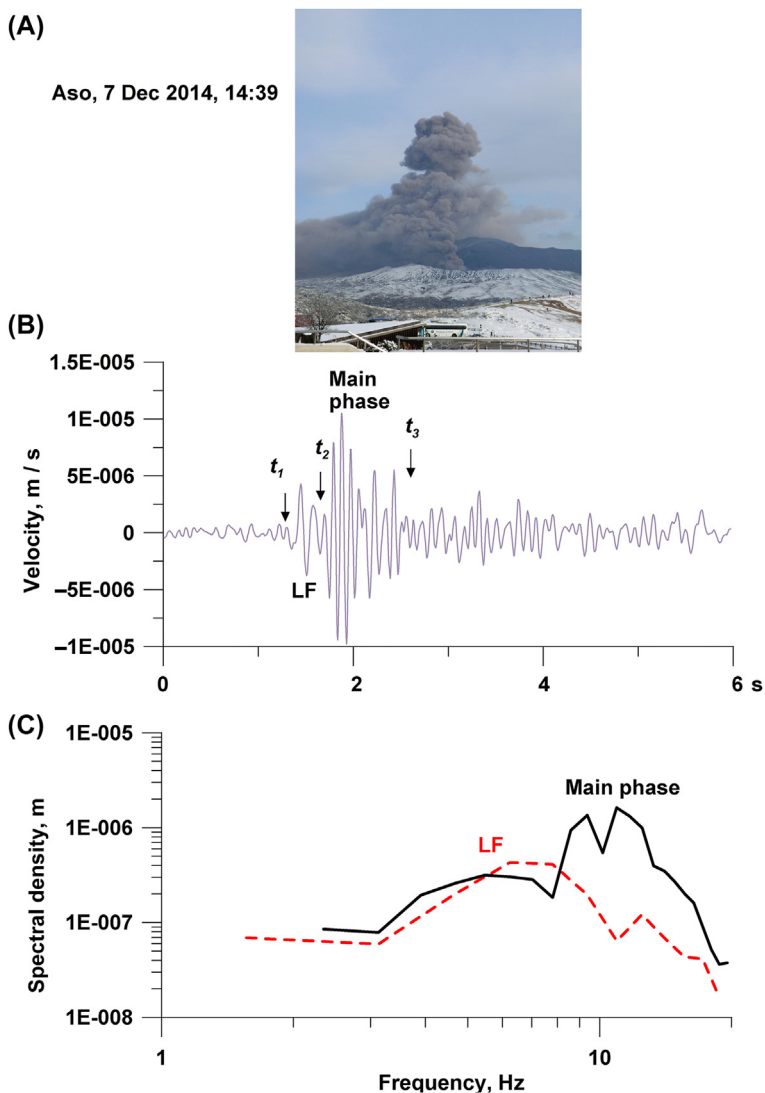


**FIGURE 14.3** Vertical component short-period velocity record of the seismic signal (B) and its Fourier spectrum (C) associated with Strombolian explosion at Aso volcano, Japan (December 7, 2014, 11:39). Photo of the explosion taken by Y. Sudo is shown in (A). Seismic station KAF was situated at a distance of 0.15 km from the rim of the crater. LF and main phase indicate the preliminary low-frequency phase and higher-frequency main phase of the seismic signal, respectively. *Courtesy: the Aso Volcano Museum.*

or HF + LF signal (Type 3). These preceding phases are well recognized on the vertical components but maybe not so well expressed on the horizontal components. Seismic signals associated with an explosion may be recorded

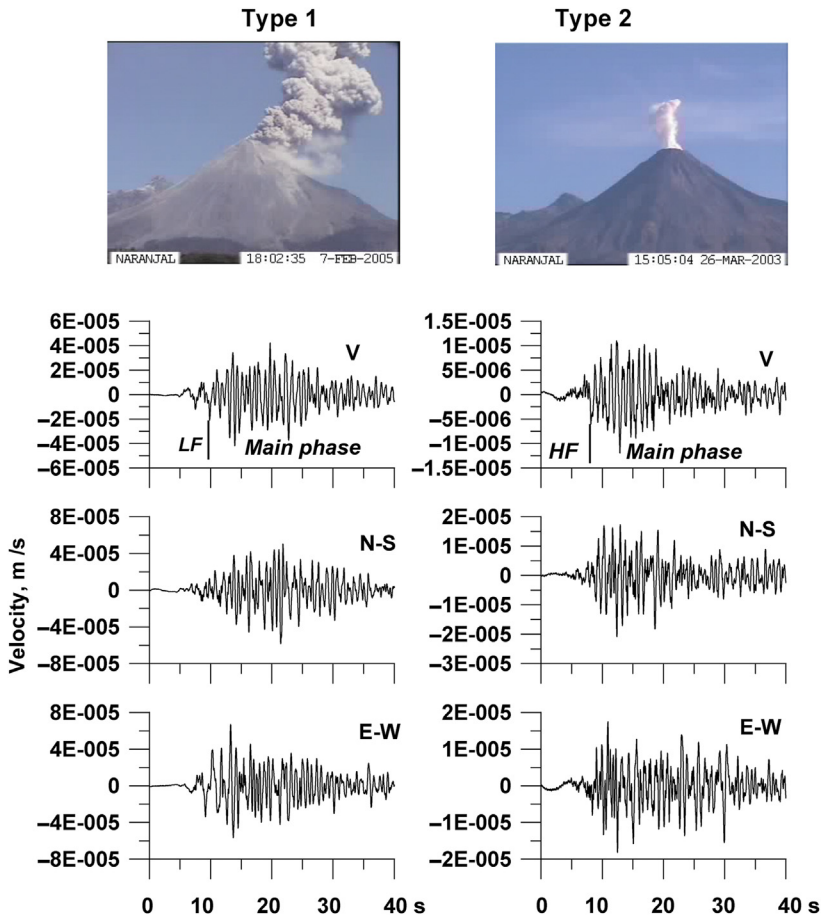


**FIGURE 14.4** Illustration of the four types of seismic signals associated with Vulcanian explosions at Volcán de Colima during the 2003–07 activity. Every panel shows the video image of the explosion (above) and three components of associated broadband seismic signals (below). Seismic station EZ5 was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 15 km S from the crater. LF and HF indicate preliminary low- and high-frequency signals, respectively.

without any preceding signal also (Type 4). Domination of the main phase on the seismic records of explosive events constrains the peak frequency between 1 and 2 Hz for all these signals (Fig. 14.5).

### 14.1.3 Plinian Explosions

The Plinian eruption (VEI 6; [Poli and Shapiro, 2022](#)) of submarine andesitic Hunga Tonga-Hunga Ha’apai Volcano (HTHH) in mid-January 2022 was

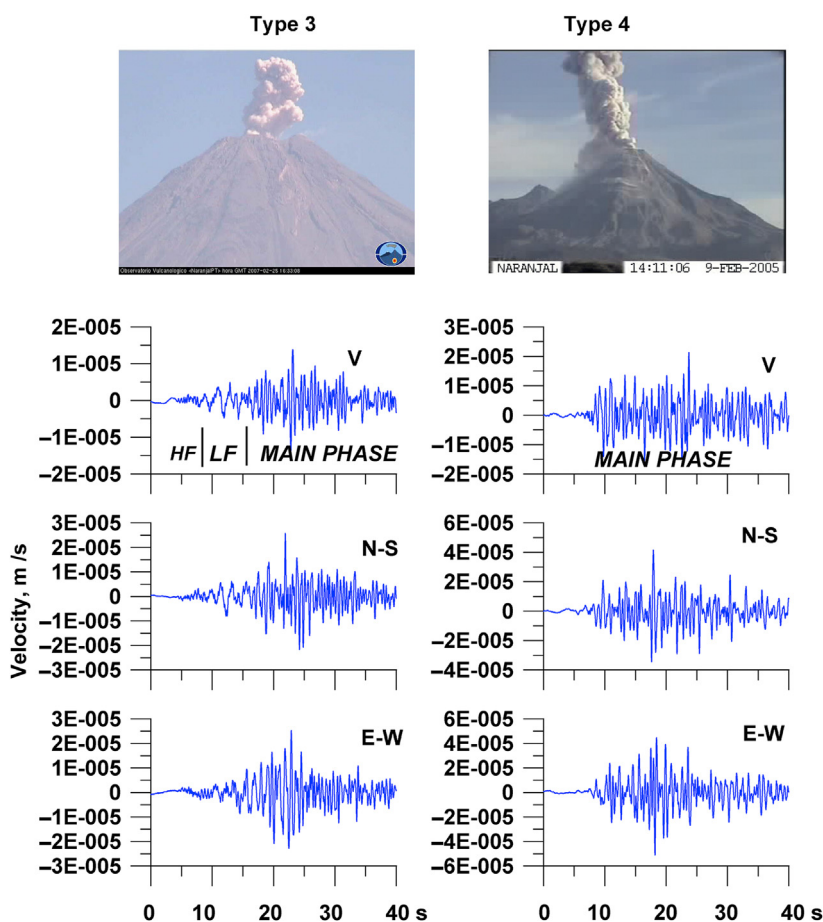


FIGURE 14.4 (Continued)

arguably the most violent volcanic eruption in the past more than 100 years. The cataclysmic explosive event onset on January 15 at 04:02 UTC (17:02 local time), with eruption intensity increasing drastically at  $\sim 04:08$  UTC and producing a plume that reached a maximum height of  $\sim 58$  km at 04:30 UTC (BGVN, 2022; Yuen et al., 2022). Fig. 14.6A demonstrates the satellite image of the eruption plume, taken at 05:00 UTC on January 15, 2022. The top of the plume measured at least 600 km in diameter (BGVN, 2022).

Yuen et al. (2022) stacked the P-wave arrivals at teleseismic distances of 70–90 degrees, where the seismic recordings were the densest and most coherent among different stations. The P-wave arrivals were picked, and traces were aligned and stacked according to the picked times. The stacked ground velocity and displacement seismograms (Fig. 14.6B) showed four

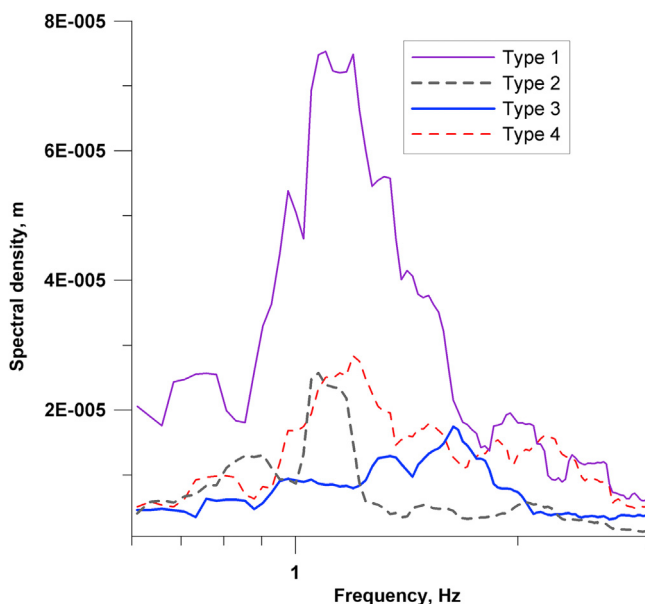


**FIGURE 14.5** Fourier spectra of four types of seismic signals (vertical component) associated with Vulcanian explosions at Volcán de Colima during the 2003–07 activity and shown in Fig. 14.4. *Courtesy: Colima Volcano Observatory.*

significant seismic events: E-1, E-2, E-3, and E-4, with similar waveforms, suggesting episodic eruptive activities. The final event (E-4) is very close to E-3 in time and is also considerably smaller in amplitude than the first three events (Yuen et al., 2022).

The moment-tensor analysis, presented by Thurin et al. (2023), has supported the representation of the HTHH eruption as a multi-event explosive seismic source, represented by at least four shallow explosions, spanning  $\sim 270$  seconds, and having a cumulative moment magnitude of  $M_w$  6.3.

#### 14.1.4 Phreatomagmatic Explosions

Fig. 14.7 shows seismic waveforms of phreatomagmatic explosion recorded during the lateral 2000 eruption at Usu volcano (see Chapter 7) and their Fourier spectra. Broadband waveform (Fig. 14.7B), recorded at a distance of 5 km from the crater, is characterized by a sequence of HF and LF pulses supposedly associated with the sequence of frequent explosive events during a phreatomagmatic eruption. The same groups of pulses but of higher frequency content are observed on short-period seismograms (Fig. 14.7C) recorded at a distance of 21 km from the crater. The main phase, similar to that observed in the seismic records of Strombolian and Vulcanian explosions, is absent. Fourier spectra of these seismic records (Fig. 14.7D) are

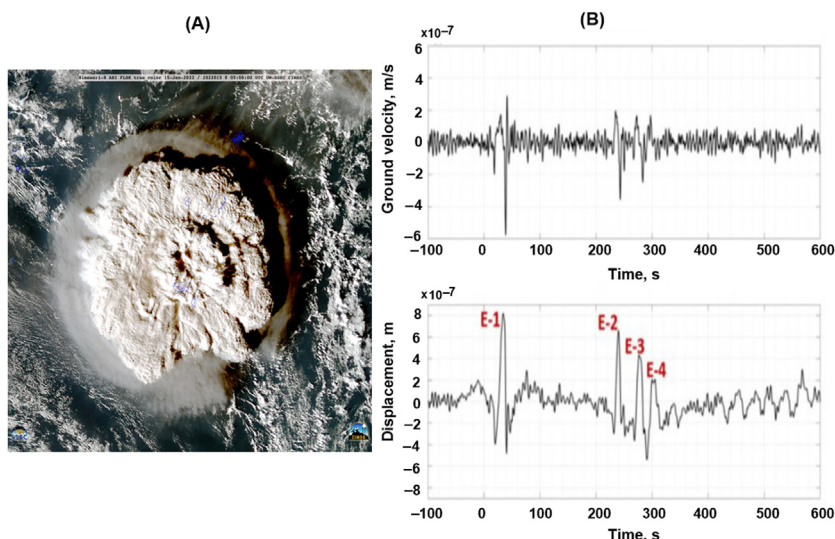


**FIGURE 14.6** (A) HIMAWARE-8 satellite image, showing the significant eruption plume, that rose from HTHH, that was taken at 05:00 UTC on January 15, 2022. Tongan islands are outlined in blue. Courtesy: CIMSS. (B) The waveforms of stacked ground velocity (above) and displacement (below) seismograms, generated by the Plinian HTHH explosive eruption. In B, four episodic seismic events, E-1 to E-4 are shown. (A) From *Global Volcanism Program, 2022. Report on Hunga Tonga-Hunga Ha'apai (Tonga)* (Bennis, K.L., and Venzke, E., eds.). *Bulletin of the Global Volcanism Network*, 47:3. Smithsonian Institution; (B) From Yuen, D.A., Melissa A. Scruggs, M.A., Frank J. Spera, F.J. et al. (2022). *Under the surface: Pressure-induced planetary-scale waves, volcanic lightning, and gaseous clouds caused by the submarine eruption of Hunga Tonga-Hunga Ha'apai volcano. Earthquake Research Advances*, 2, 100134.

very similar for the interval between 0.1 and 5 Hz and represent a stable amplitude level within this spectral interval.

Similar broadband waveforms were reported for phreatic explosions at Aso volcano, Japan, during the Aso94 experiment (Kawakatsu et al., 2000).

## 14.2 NATURE OF THE SEISMIC SIGNALS OF STROMBOLIAN AND VULCANIAN EXPLOSION EARTHQUAKES

### 14.2.1 Comparison of the Contemporary Video and Seismic Records During an Explosion

#### 14.2.1.1 Strombolian Explosions

The contemporary records of the seismic and acoustic signals at Stromboli during its explosive activity from March 1962 to April 1962, performed by Peterschmitt and Tazieff (1962), showed that the initial seismic vibrations had appeared 12–23 seconds before the visible eruption. As Chouet et al. (1997) stated, the onset of the higher frequency signals, recorded at

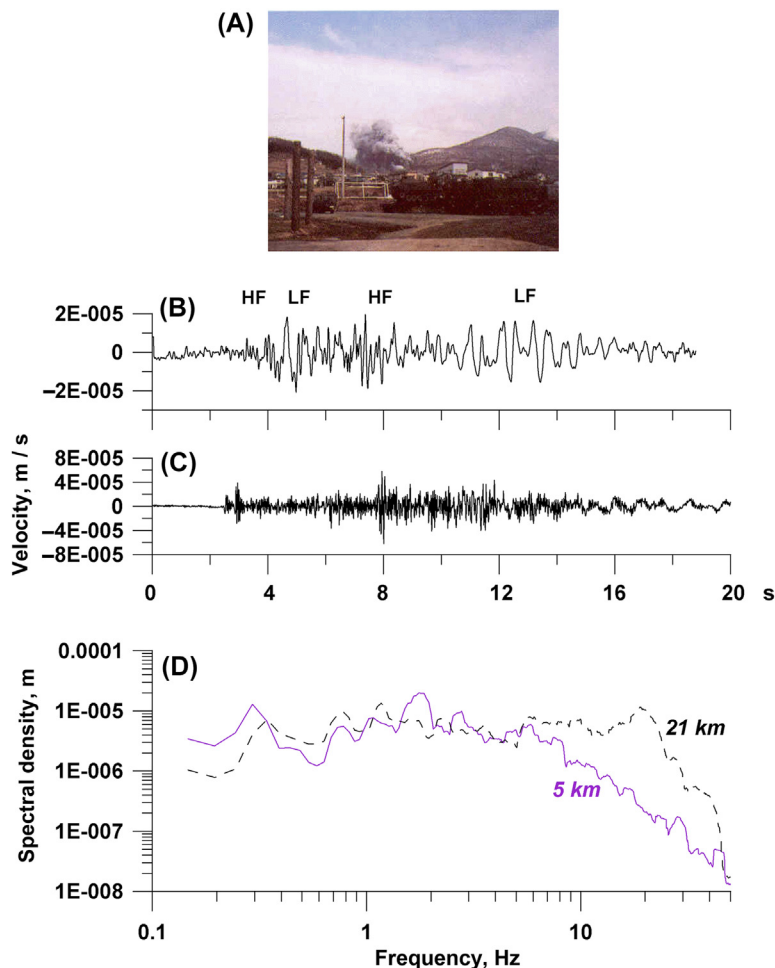


**FIGURE 14.7** Video image (A), broadband seismic record (velocity, vertical component) (B), and short-period seismic record (velocity, vertical component) (C) associated with the March 31, 2000 (13:08) phreatomagmatic explosion at Usu volcano (lateral cone Nishiyama). Broadband seismic station was situated at a distance of 5 km from the crater; short-period station, at a distance of 21 km from the crater. (D) Fourier spectra of seismic records of this explosion recorded at a distance of 5 km (broadband, *solid line*) and 21 km (short period, *broken line*). LF and HF indicate low- and high-frequency signals, respectively. Video image was taken from JMA, 2003. *Report on the Eruption of Usuzan Volcano in 2000. Technical Report of Japan Meteorological Agency 124. JMA, Tokyo (in Japanese, English abstract).* Broadband seismic record is a courtesy of Tohoku University; short-period seismic record is a courtesy of Japan Meteorological Survey.

Stromboli volcano in 1992, was noted to coincide with the visually observed onset of the eruptive jet, while no visible surface activity could be linked with the precursory LF onset. The contemporary seismic-video records of the explosion process allow working out these observations in detail.



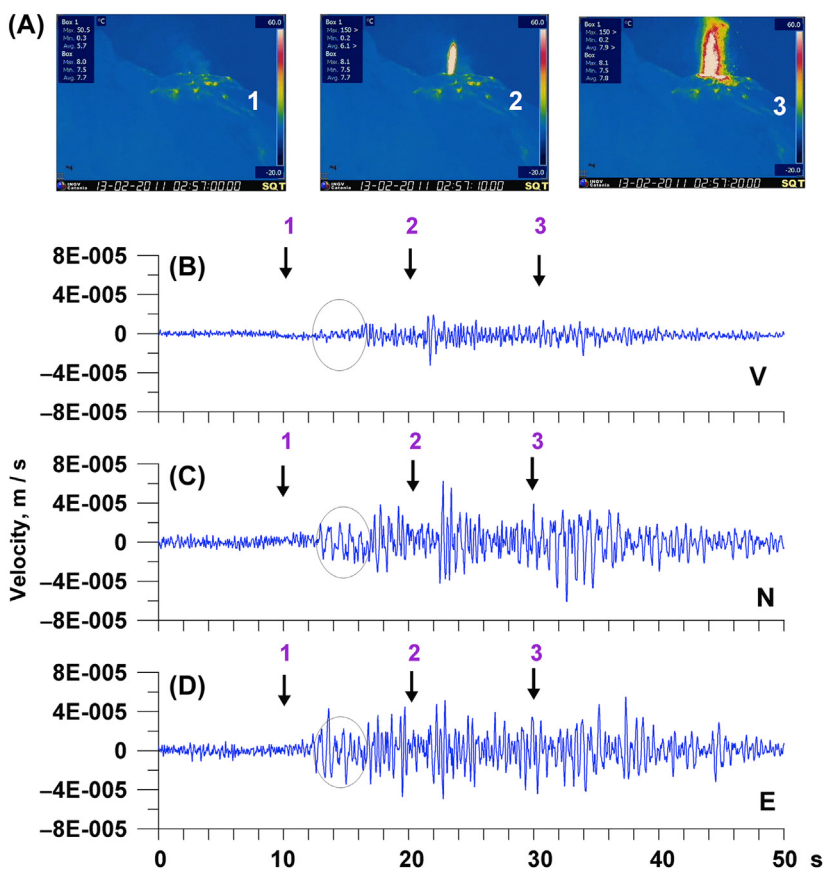


**FIGURE 14.8** Contemporary video (A) and three-component broadband velocity seismic records (B–D) during the Strombolian explosion (February 13, 2011, 02:57) at Stromboli volcano. Seismic station was situated at a distance of 0.5 km from the crater. Video station was situated at a distance of about 1 km from the crater. LF and HF indicate low- and high-frequency signals, respectively. 1, 2, and 3 correspond to the timing of images and the seismic record. *Courtesy: Istituto Nazionale di Geofisica e Vulcanologia, Sezione di Catania.*

Fig. 14.8 shows a three-component broadband seismic record of an explosion, which occurred at Stromboli volcano in January 2011, together with the contemporary video record. It is seen that the initial LF pulse of the seismic signal (better seen in horizontal components) is recorded before the outlet of the volcanic products from the crater and supposedly corresponds to the process occurring within the conduit before the explosion. The HF and larger amplitude signals of the main phase, which arrived 5 seconds later, were recorded just before and during the outlet of volcanic products from the crater.



14.2.1.2 *Vulcanian Explosions*

Fig. 14.9 shows the contemporary records of broadband seismic signal and video images of the Vulcanian explosion (Type 3) at Volcán de Colima, Mexico. Type 3 of seismic signals associated with Vulcanian explosions was selected because contains both preliminary phases, HF and LF. It is seen that the first superficial manifestation of the explosion, marked by an arrow in image 2, appeared 26 seconds after the onset of the seismic signal. Therefore the initial part of the seismic record, containing the preliminary phases LF and HF, may be considered to be generated by the process occurring within the volcano conduit before an explosion.

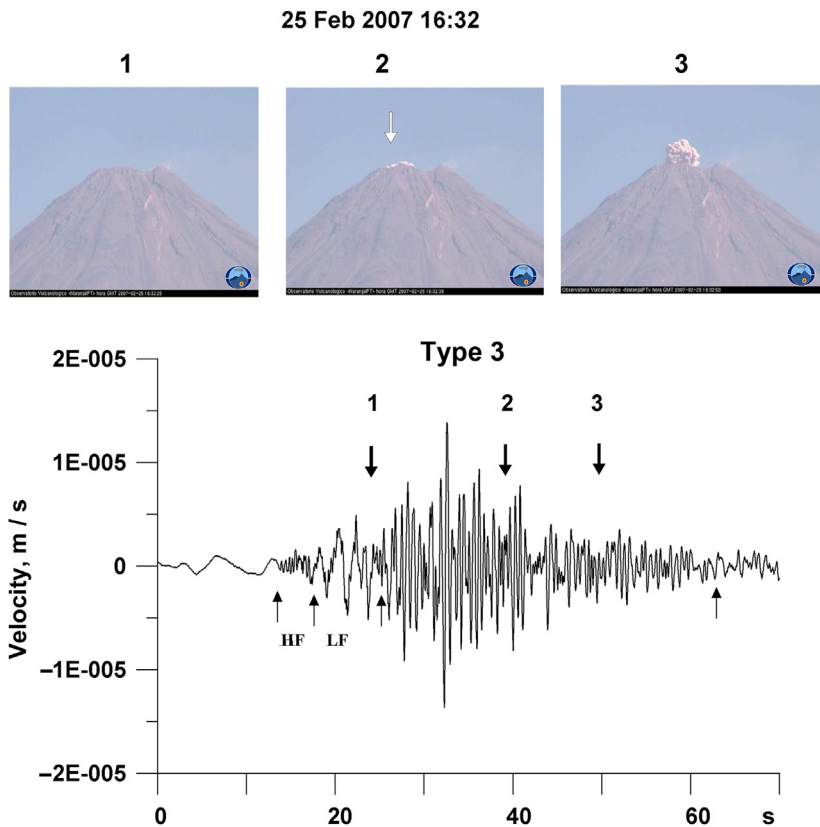


**FIGURE 14.9** Contemporary video and broadband velocity seismic records (vertical component) during the Vulcanian explosion at Volcán de Colima (February 22, 2007, 16:32). Seismic station was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 15 km S from the crater. LF and HF indicate low- and high-frequency signals, respectively. 1, 2, and 3 correspond to the timing of images and the seismic record. *Courtesy: Colima Volcano Observatory.*

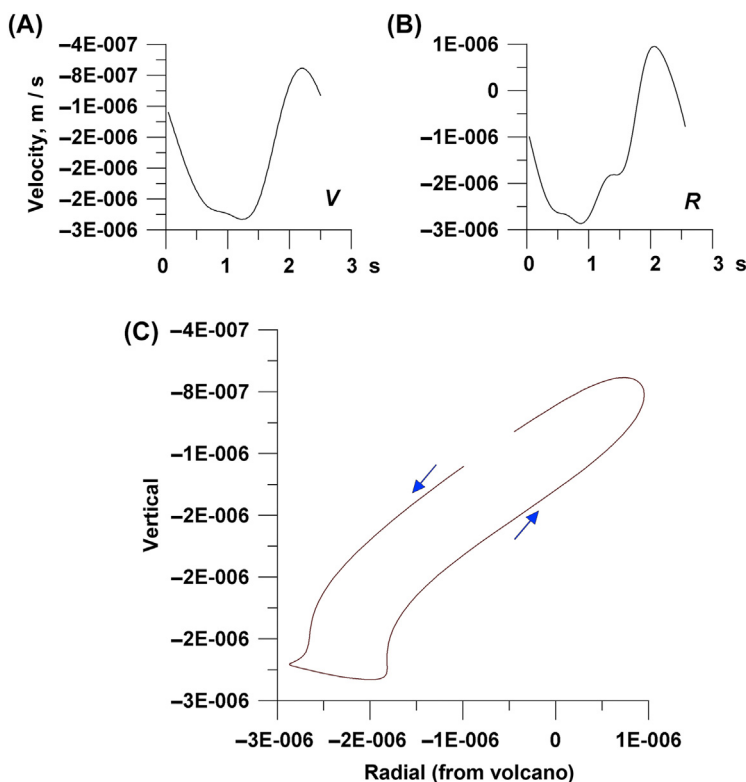


**FIGURE 14.10** Particle motions of preliminary low-frequency signal of seismic records associated with Strombolian volcanic explosion shown in Fig. 14.7 The seismic records [(A) vertical component and (B) radial component] were low-pass filtered at 0.5 Hz. (C) The trajectory of retrograde particle motion. Courtesy: Istituto Nazionale di Geofisica e Vulcanologia, Sezione di Catania.

## 14.2.2 Type of Waves Composing the Seismic Signal of an Explosion

### 14.2.2.1 Strombolian Explosions

Fig. 14.10 shows the particle motions for the preliminary LF signal of the seismic record associated with the explosion at Stromboli volcano (see Fig. 14.7). To avoid the influence of HF noise, the records of vertical and radial components were low-pass filtered at 0.5 Hz. It displays a retrograde elliptical particle motion indicating a Rayleigh wave. The same retrograde elliptical particle motion was shown by Rowe et al. (1998) for the initial 15-s signals of broadband seismic records of explosive events at the lava lake of Erebus volcano.

### 14.2.2.2 Vulcanian Explosions

Fig. 14.11 shows the particle motion for the preliminary LF signal of the seismic record associated with the explosion at Volcán de Colima. To avoid

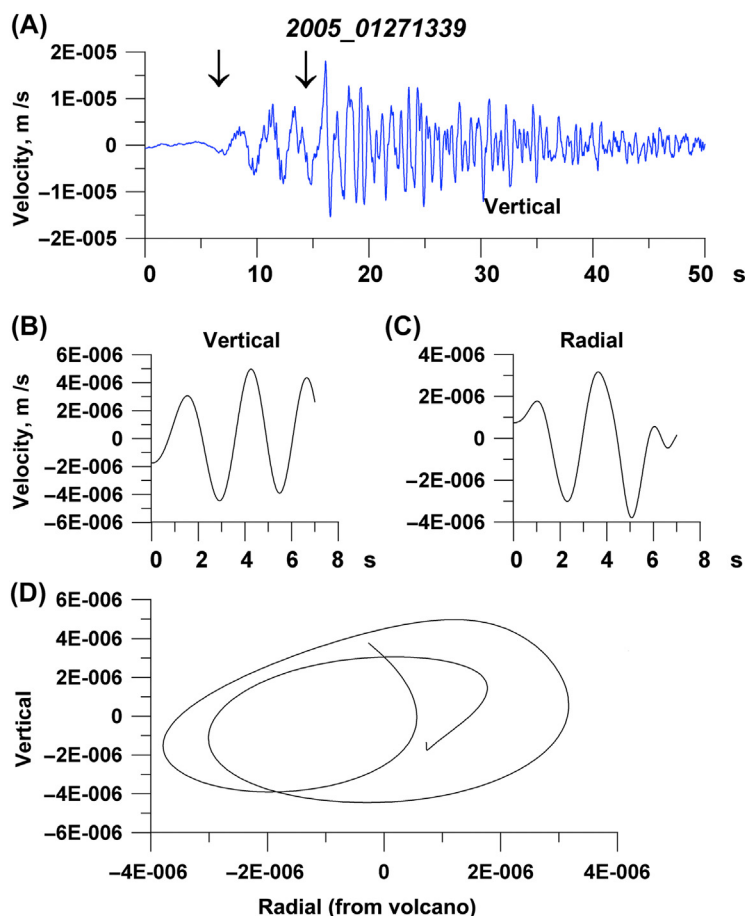


**FIGURE 14.11** Particle motions in the seismic record of preliminary low-frequency signal (LF) of earthquake associated with Vulcanian explosion at Volcán de Colima (January 27, 2005, 13:39). (A) The seismogram (velocity, vertical component); the LF signal is marked by arrows. (B, C) The vertical and radial components of the LF impulse were low-pass filtered at 0.5 Hz. (D) The trajectory of retrograde particle motion. The seismic station was situated at a distance of 4 km from the crater. From Zobin, V.M., Navarro, C., Reyes-Dávila, G., Orozco, J., Bretón, M., Tellez, A., Reyes-Alfaro, G., Vázquez, H., 2006b. The methodology of quantification of volcanic explosions from broadband seismic signals and its application to the 2004–2005 explosions at Volcán de Colima, México. *Geophysical Journal International* 167, 467–478.

the influence of HF noise, the records of vertical and radial components were low-pass filtered at 0.5 Hz. The record is short but it is possible to see that the retrograde motion of particles forms well-expressed ellipses in the vertical plane that is characteristic of Rayleigh waves (Zobin et al., 2006a,b). Tameguri et al. (2002) studied the particle motions of Sakurajima explosions derived from broadband records. They also obtained that the low-frequency

phase (LP equivalent to LF in Fig. 14.4) shows an elliptical orbit and a retrograde rotation in a vertical cross-section along the direction of wave propagation. The same result about the Rayleigh nature of the preliminary LF phase was obtained by [Zobin and Martínez \(2010\)](#) for the seismic records associated with Vulcanian explosions of the Popocatepetl volcano.

[Tameguri et al. \(2002\)](#) studied the wave nature of preliminary HF signals recorded by short-period borehole seismometers during the 1999 Sakurajima Vulcanian explosions and concluded that they are P-waves.

### 14.3 SOURCES OF STROMBOLIAN AND VULCANIAN EXPLOSION EARTHQUAKES AND THEIR QUANTIFICATION

The complex structure of the seismic signals associated with volcanic explosions as well as the problem of identification of the nature of waveforms, composing these seismic signals, and unclear arrivals of the seismic phases make it difficult to study the source process of explosion earthquakes. To resolve the problem, many research scientists study the source of explosion events as a source that generates long-period and very-long-period seismic signals neglecting short-period components. The bandpass filters the original broadband seismic records leaving only long-period and very-long-period vibrations (as shown in Fig. 14.12) and then performs the waveform inversion obtaining the characteristics of a centroid as the source of these vibrations (e.g., [Arciniega-Ceballos et al., 1999](#); [Chouet et al., 2003, 2005](#)). This approximation gives the integral characteristics of the source of explosion events neglecting the details of the process.

The results that were obtained using long-period and very-long-period approximations of the seismic signals associated with volcanic explosions are discussed in Chapter 15 together with other applications of long-period signals generated by volcanic processes. In this chapter, the author discusses the results that were obtained using the fine structure of the seismic signal as the information about the development of ascending magmatic process within the conduit before an explosion and during the explosion (e.g., [Tameguri et al., 2002](#); [Zobin et al., 2006b, 2009a,b](#)).

[Tameguri et al. \(2002\)](#) discriminated four phases of the seismic records of Vulcanian explosions at the andesitic Sakurajima volcano and estimated the moment tensors for the sources generated in these phases. [Zobin et al. \(2006a,b, 2009a,b\)](#) developed a two-stage model of the explosive process for explosions at Volcán de Colima and Popocatepetl which allows quantifying the volcanic explosions in the units of force (N) and energy (J) from unfiltered broadband seismic records (see Chapter 11). [Kumagai et al. \(2011\)](#) proposed the ascending seismic source for the explosion at Tungurahua volcano, Ecuador, considering that a pressure wave propagates along the magma conduit triggering fragmentation of magma at shallow depths.

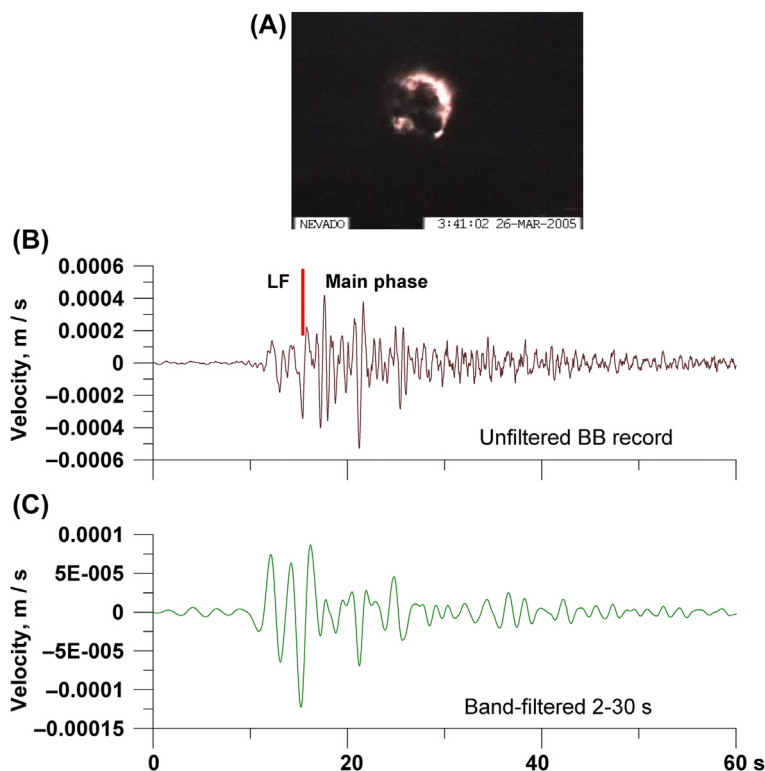


**FIGURE 14.12** Broadband seismic record (velocity, vertical component) and video image (A) of Vulcanian explosion at Volcán de Colima (March 23, 2005, 03:41). (B) Unfiltered record and (C) the same record but bandpass filtered at 2–30 s. The seismic station was situated at a distance of 4 km from the crater. Video station was situated at a distance of about 5 km N from the crater. LF indicates a preliminary low-frequency signal. *Courtesy: Colima Volcano Observatory.*

### 14.3.1 Multiple Source of Explosions

Tameguri et al. (2002) discussed the nature of broadband seismic records of Vulcanian explosions at Sakurajima volcano occurring from August 1999 to December 1999. Fig. 14.13 shows a broadband seismic record of the 1999 Sakurajima explosion obtained on the seismic station situated at a distance of 2.75 km. This seismic record corresponds to Type 3 of Vulcanian events shown in Section 14.1.2. Tameguri et al. (2002) analyzed three phases: initial phase P and the second phase D within the first group of short-period pulses (or phase HF in Fig. 14.4), and the phase LP corresponding to the group of long-period pulses (or phase LF in Fig. 14.4). The phases P and D were identified as P-waves, and LP phase as composed of Rayleigh waves.

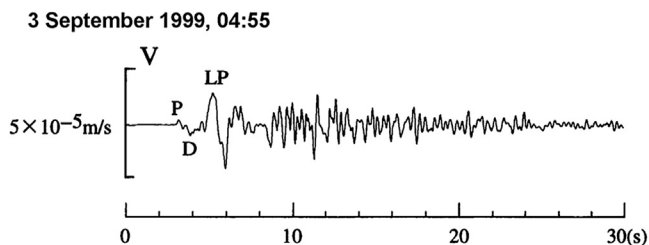


**FIGURE 14.13** Broadband seismic signal (velocity, vertical component), associated with Vulcanian explosion at Sakurajima volcano, Japan, and recorded at a distance of 2.8 km from the crater. P, D, and LP are three phases. After Tameguri, T., Iguchi, M., Ishihara, K., 2002. *Mechanism of explosive eruptions from moment tensor analyses of explosion earthquakes at Sakurajima volcano, Japan. Bulletin of the Volcanological Society of Japan* 47, 197–215.

Moment tensors of the seismic sources, generated by these phases, were estimated from waveform inversions for 14 explosions. The source of phase P was characterized by three diagonal components with the same positive values for all events. It corresponds to the isotropic expansion source (Fig. 14.14). The sources of phase D (Fig. 14.13) were characterized by horizontal dipoles twice as large as the vertical which is in accordance with the theoretical ratio among the three diagonal components caused by a volume change of cylindrical source (2:2:1). The sources generated P and D phases were located at the depth of 2 km beneath the crater. LP phase was excited by two sources: the first source with an isotropic expansion and the second horizontal contraction source at shallow depths of 0.25–0.5 km beneath the crater bottom (Fig. 14.15). The values of seismic moments obtained for the sources varied from  $10^{11}$  to  $10^{12}$  N m.

Based on the conceptual model of the Sakurajima explosions, proposed by Ishihara (1990), the sequence of seismic sources was supposed (Tameguri et al., 2002) to correspond to the following eruptive process (Fig. 14.16). The gas pocket, created by the exsolution of volatiles in the upper part of the shallow magma chamber at the depth of about 2 km (phases P and D), abruptly moves up through the conduit to the surface generating the phase LP. Then the explosion, accompanied by the outflow of gas, ash, and volcanic blocks, occurs (the coda of the LP signal and the second HF group of the seismic record). Therefore the explosion occurs after the radiation of the P and D phases and during the generation of the LP phase.

### 14.3.2 Two-Stage Conceptual Models of Explosive Process

The two-phase structure of the seismic signals associated with Strombolian and Vulcanian explosions allows considering that these two components of the seismic radiation are linked to different source processes that are generated by the explosions that were proposed by Martini et al. (2007) for

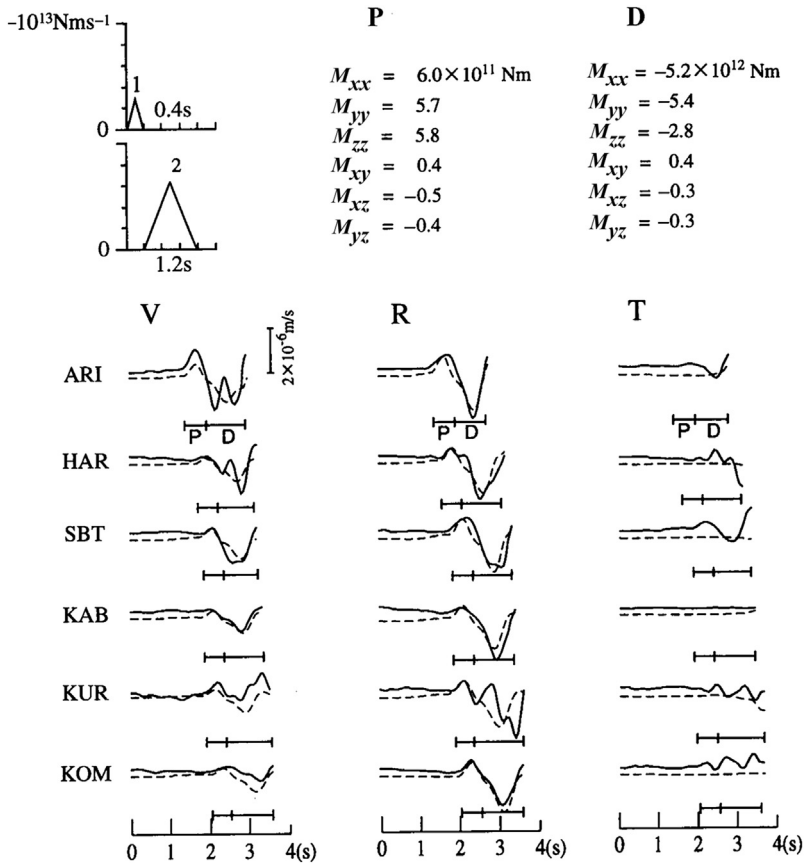


**FIGURE 14.14** Broadband waveform inversion of the P and D phases. The explosion earthquake occurred on September 15, 1999, 04:14. Solid and broken traces indicate observed and synthetic waveforms, respectively. V, R, and T indicate vertical, radial, and tangential components, respectively. Two triangular source time functions 1 and 2 corresponding to the P phase and D phase are displayed in the upper left and obtained moment-tensor components are shown on the right side. Bars with symbols “P” and “D” under the waveforms represent time windows of waveform inversion. The observed and synthetic waveforms were low-pass filtered at 5 Hz. From Tameguri, T., Iguchi, M., Ishihara, K., 2002. Mechanism of explosive eruptions from moment tensor analyses of explosion earthquakes at Sakurajima volcano, Japan. *Bulletin of the Volcanological Society of Japan* 47, 197–215.

Strombolian explosions at Stromboli. Generally, the conceptual models of the explosive process, based on contemporary seismic and video observations (see Section 14.2.1), may consist of two stages: preexplosive ascent of gas or magma within the volcanic conduit and following explosion. The processes of the preexplosive ascent of gas or magma within the volcanic conduit generate the initial seismic phases and the coexplosive subsurface processes generate the main phases of the seismic signals.

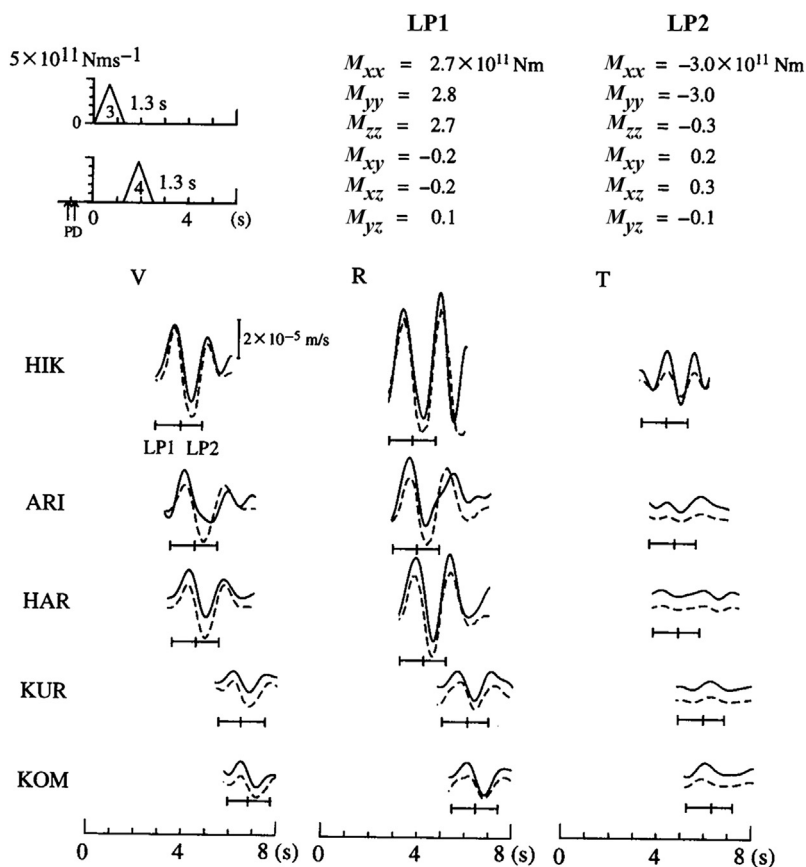


**FIGURE 14.15** Broadband waveform inversion of preliminary LP phase. The explosion earthquake occurred on August 27, 1999, at 19:21. Solid and broken traces indicate observed and synthetic waveforms, respectively. V, R, and T indicate vertical, radial, and tangential components, respectively. Two triangular source time functions 3 and 4 corresponding to the LP1 phase and LP2 phase are displayed in the upper left and obtained moment-tensor components are shown on the right side. Bars with symbols "LP1" and "LP2" under the waveforms represent time windows of waveform inversion. The observed and synthetic waveforms were low-pass filtered at 1 Hz. From Tameguri, T., Iguchi, M., Ishihara, K., 2002. Mechanism of explosive eruptions from moment tensor analyses of explosion earthquakes at Sakurajima volcano, Japan. *Bulletin of the Volcanological Society of Japan* 47, 197–215.

### 14.3.2.1 Strombolian Explosions

Strombolian eruptions are mildly explosive events that are driven by a large bubble of magmatic slug rising up the conduit and bursting at the surface. A slug represents a particular type of gas bubble flowing through a liquid-filled pipe (Taddeucci et al., 2015). Therefore the following conceptual two-stage model may be proposed: the initial phase of the seismic signal corresponds



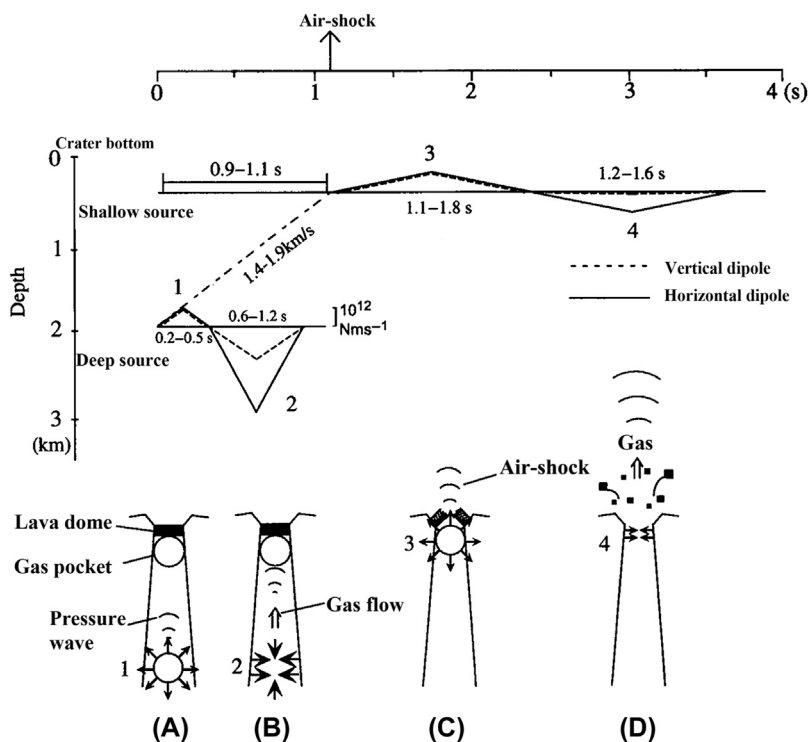


**FIGURE 14.16** Time sequence of origin times and depths of four sources generating P, D, and LP phases of explosion earthquake. Solid and broken lines indicate source time functions of horizontal and vertical dipoles, respectively. Mechanics of explosive eruption from seismic data is schematically shown in the lower part. (A) An isotropic expansion occurs at a depth of 2 km beneath the crater. Then, a pressure wave generates and propagates through the conduit up to the crater. (B) A cylindrical contraction is generated by the pressure decrease of the source region due to escape gases upward in the conduit. (C) Outbreak of gas pocket beneath a lava dome is started by shallow expansion, which is excited by a pressure wave propagating from the deep source. At the same time, air shock is ejected from the crater bottom. (D) Gases are released from the uppermost part of the conduit. Horizontal contraction starts 1–2 s after the beginning expansion. From Tameguri, T., Iguchi, M., Ishihara, K., 2002. Mechanism of explosive eruptions from moment tensor analyses of explosion earthquakes at Sakurajima volcano, Japan. *Bulletin of the Volcanological Society of Japan* 47, 197–215.

to the process of ascent of the gas slug to the surface, and the main phase of the signal is generated by the explosion at the surface. The seismic energy of the initial phase may be used as a measure of a force governing the movement of the gas slug while the seismic energy of the main phase may be used as a parameter proportional to the energy of an explosion. The time delay between the arrivals of the two phases is supposed as a function of the initial depth of the slug ascent.

### 14.3.2.2 Vulcanian Explosions

As shown in Section 14.1.2, the three types of two-phase seismic signals associated with Vulcanian explosions were defined. They differ in the waveforms of the initial phases, LF for Type 1, HF for Type 2, and HF + LF for Type 3. Fig. 14.17 demonstrates the conceptual model proposed by Zobin et al. (2015) to explain the nature of the three types of seismic signals. This model develops the earlier two-stage model proposed for the Type 1 Vulcanian explosion and is shown in Section 11.3.2.1 (Image 11.15 from Chapter 11).

According to the model, the process, generating the seismic signal, associated with the Type 1 Vulcanian explosions, corresponds to the magma bubbles' ascent from the fragmentation zone within the volcanic conduit due to a sudden pressure drop at the base of the supersaturated magma. The pressure disturbances in the fragmented magma-excited resonances of the volcanic conduit produce LF vibrations. The conduit overpressure, exceeding the strength of the cap of degassed crystal-rich magma, triggers the explosion at the surface (stage 2). During stage 2, the main phase of the signal is generated.

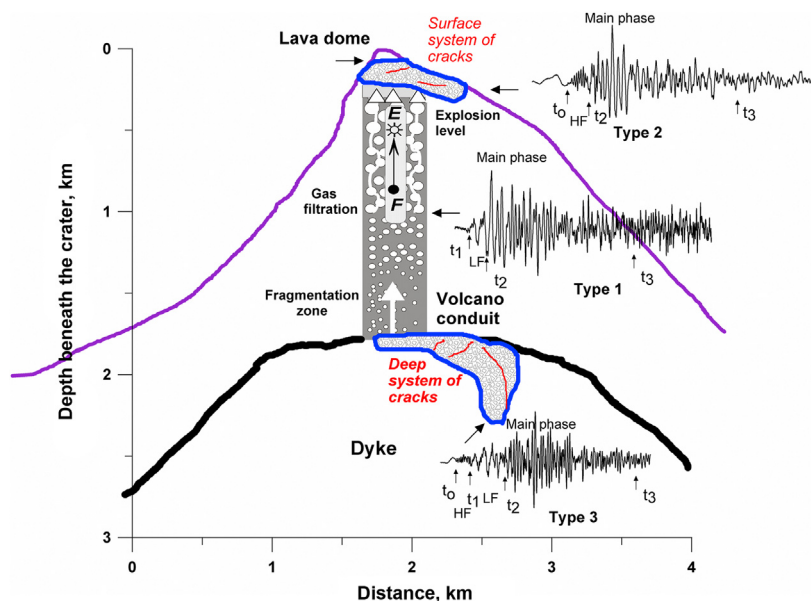


**FIGURE 14.17** Conceptual model of three types of Vulcanian explosions. Three types of seismic signals associated with explosive events are shown. The high-frequency (HF) and low-frequency (LF) initial phases and the main phase of the seismic signal are indicated. Symbols  $t_0$ ,  $t_1$ ,  $t_2$ , and  $t_3$  show the arrival and termination of the phases.  $F$  indicates the force governing the movement of the magma jet, and  $E$  indicates the site of an explosion at the explosion level. From Zobin, V.M., Peral, J.J., Nava, F., Bretón, M., 2015. Vulcanian explosions in the process of building-destruction of the lava dome of andesitic volcano: insight from the seismic signals recorded at Volcán de Colima, México. *Physics of the Earth and Planetary Interiors* 244, 1–10.

The process, generating the Type 2 Vulcanian explosions and associated seismic signals, is supposed to consist of the fracturing of the subsurface lava cap (stage 1) and a subsequent explosion (stage 2). During stage 1, the gas mixture that is collected in the sublava cap cavities near the crater reaches high-pressure levels that lead to the crack opening and propagation toward the dome border generating the HF phase. During stage 2, the main phase of the signal is generated.

The process, generating the Type 3 Vulcanian explosions and corresponding seismic signals, is proposed to be more complicated and consists of the fracturing of the magma dyke beneath the conduit with the following ascent of fragmented magma to the surface (stage 1) and a subsequent explosion (stage 2). During stage 1, the magma enters into the conduit through the deep system of cracks in the dyke at the level of the fragmentation zone generating the HF phase and triggering the magma bubbles ascent from the fragmentation zone within the volcanic conduit. The pressure disturbances in the fragmented magma-excited resonances of the volcanic conduit produce LF vibrations. The conduit overpressure, exceeding the strength of the cap of degassed crystal-rich magma, triggers the explosion at the surface. During stage 2, the main phase of the signal is generated.

### 14.3.3 Comparison of the Source Properties of Strombolian and Vulcanian Explosions

As was mentioned in [Section 14.1.1.1](#), the Vulcanian explosions, characterized by VEI 2–4, are next in the hierarchy of volcanic eruptions to Strombolian explosions, occurring with VEI 1–2. The analysis of infrasonic and thermal records for Strombolian (Stromboli and Villarrica volcanoes) and Vulcanian (Santiaguito and Fuego volcanoes) explosions, performed by [Marchetti et al. \(2009\)](#), showed that the Strombolian explosions shared low values of buoyancy, pointing to a gas-thrust dominated emission, efficient coupling of the infrasonic source to the atmosphere, and corresponded to source process with generation of a plume rich in coarse fragments. In contrast, Vulcanian-type explosions shared large buoyancy and were consistent with events in which a large part of infrasonic energy was spent in magma fragmentation leading to ascent of buoyant, ash-rich plumes. Thermal radiation energy estimated for explosions at Stromboli and Villarrica ( $10^4$ – $10^7$  J) is lower than that for events at Santiaguito and Fuego ( $10^8$ – $10^9$  J), but infrasonic energies overlap. [Marchetti et al. \(2009\)](#) demonstrated that the comparison between thermal radiation and infrasonic energies of explosions allowed quantitative discrimination between Strombolian- and Vulcanian-type explosions.

The comparative analysis of seismic signals associated with Strombolian- and Vulcanian-type explosions gives an additional base for quantitative discrimination between the two types of explosions. [Fig. 14.18](#) demonstrates the waveforms (A, B) and corresponding Fourier spectra (C, D) of the seismic signals recorded during three Strombolian explosions at Aso volcano and three Vulcanian Type 1 explosions at Volcán de Colima. They display the

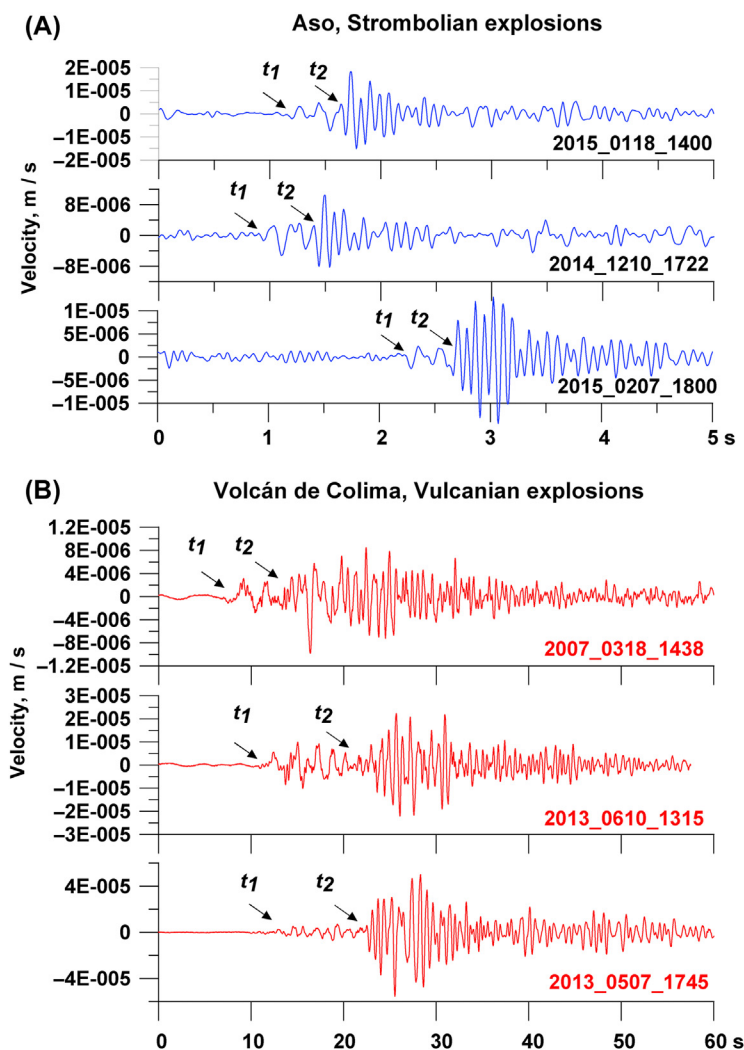


**FIGURE 14.18** Waveforms (A, B) and corresponding Fourier spectra (C, D) of two-phase seismic signals recorded during (A) three Strombolian explosions at Aso volcano (short-period seismic station KAF, vertical component) and (B) three Vulcanian explosions at Volcán de Colima (broadband seismic signals, vertical component, 4 km from the crater) Arrows labeled with  $t_1$  and  $t_2$  indicate the arrivals of two phases. The dates of seismic events are marked as YYYY\_MMDD\_HHMM.

similar two-phase structure of the signals with the lower frequency initial phases and subsequent higher frequency, higher amplitude main phases (Fig. 14.18A and B). At the same time, the durations of both phases are significantly larger for Vulcanian explosions (within 40–45 seconds) compared

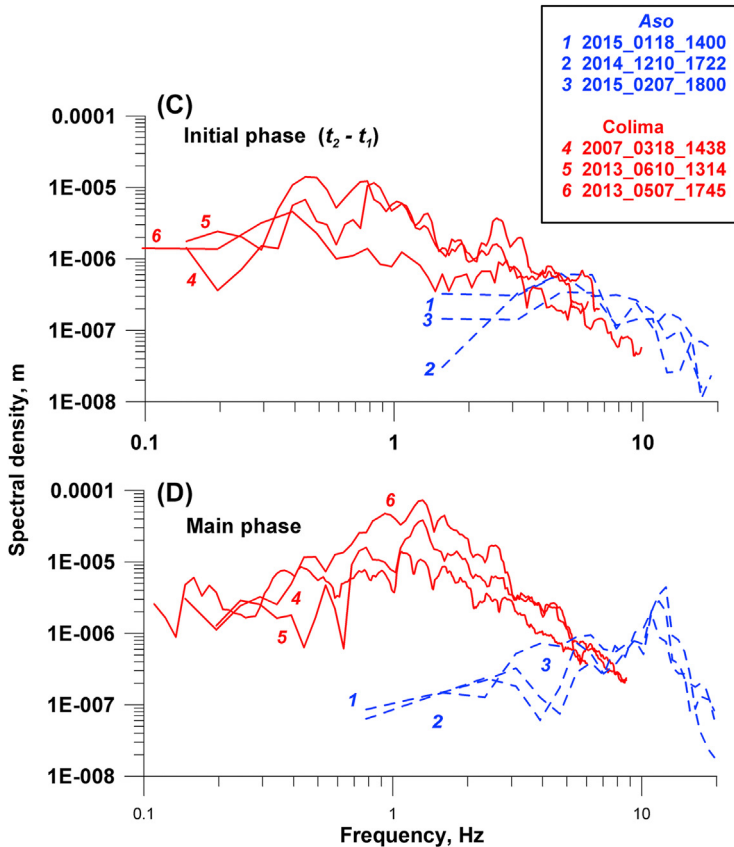


FIGURE 14.18 (Continued)

to Strombolian explosions (within 2.5–3 seconds). A great difference is seen in comparison of Fourier spectra (Fig. 14.18C and D). The initial phases of Vulcanian signals are characterized by the peak frequencies within 0.4–0.8 Hz, while the initial phases of Strombolian signals are characterized by the peak frequencies within 4–6 Hz. The peak frequencies of the main phase spectra varied within 1.0–1.3 Hz for Vulcanian signals and within 11–13 Hz for Strombolian signals.

### 14.3.4 Source Scaling of the Seismic Signals Associated With Vulcanian and Strombolian Explosions

#### 14.3.4.1 Vulcanian Eruptions

The source parameters, obtained for 120 Colima explosions and 26 Popocatepetl explosions of Type 1 (Zobin et al., 2009b), varied over the following ranges:

the time of the movement of fragmented magma in the conduit before the explosion varied from 1 to 13 seconds; the counterforce of the eruption,  $F$ , acting in the conduit before an explosion and governing the fragmented magma movement to the surface spanned three orders of magnitude ranging from  $8.3 \times 10^7$  to  $8.5 \times 10^{10}$  N; and the energy,  $E$ , of explosion spanned six orders of magnitude ranging from  $4.6 \times 10^7$  to  $1.5 \times 10^{13}$  J.

The first scaling law for counterforce of explosion earthquakes was proposed by Nishimura and Hamaguchi (1993). They obtained the relationship between the peak amplitude of the single counterforce  $F$  and the source duration  $\tau$  of the single force using the seismic records of large volcanic explosions of the world. The obtained forces varied from  $2.6 \times 10^{12}$  N for the great May 18, 1980, explosion at Mount St. Helens, Cascades (Kanamori et al., 1984) to  $10^7$  N for small explosions at Mount Tokachi, Japan (Nishimura, 1995). For large events ( $F > 10^9$  N), the empirical law

$$\log F(N) = 2\log \tau(s) + 9.6 \quad (14.1)$$

was obtained (Nishimura and Hamaguchi, 1993; Fig. 14.19A). For smaller events, the dependence of the peak force upon the source duration is too weak;  $\tau$  was almost a constant value. Nishimura (1995) considered that this difference was caused by the difference in a “critical magma radius” of the hypothetical spherical gas-rich magma batch that produced the volcanic explosion.

The results of Zobin et al. (2009b), obtained for the Volcán de Colima and Popocatepetl explosions, support the scaling law of Nishimura and Hamaguchi having the same tendency of  $F \sim \tau^2$ . Two types of scaling were observed again (Fig. 14.19B). The first group joins the large events with a counterforce greater than  $4 \times 10^9$  J; the second group consists of smaller events. The large events of both Colima and Popocatepetl explosions were correlated with the source duration with the coefficient of correlation  $R = 0.64$  ( $R_{\text{crit}99\%} = 0.44$ ) and their regression equation was

$$\log F = 1.59\log \tau + 9.90 \quad (14.2)$$

The group of small events in Zobin et al. (2009b) was more numerous than the group of small events shown by Nishimura (1995) with a greater difference in  $\tau$ . It allowed Zobin et al. (2009b) to calculate the regression line for small events also:

$$\log F = 1.77\log \tau + 8.31 \quad (14.3)$$

with the coefficient of correlation  $R = 0.60$  ( $R_{\text{crit}99\%} = 0.25$ ).

The regression between the energy of explosions,  $E$ , and the counterforce of the eruption,  $F$ , obtained for the explosions occurring on Volcán de Colima and Popocatepetl (Zobin et al., 2009b)

$$\log E = 1.87\log F - 7.29 \quad (14.4)$$

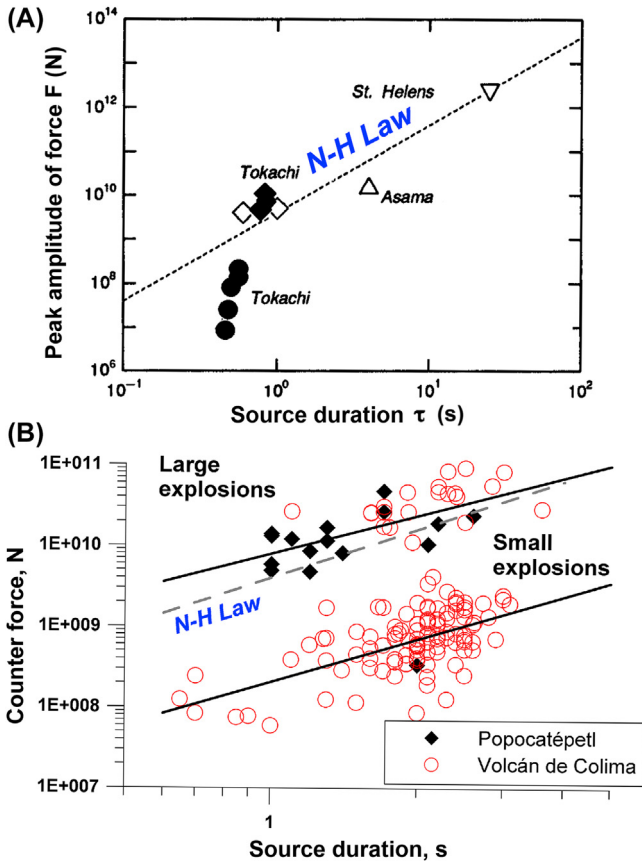


**FIGURE 14.19** Relationship between the counterforce of the eruption and the source duration of a single force. (A) Scaling law obtained by Nishimura and Hamaguchi (1993) for different volcanoes of the world; (B) relationship for the explosion earthquakes of Volcán de Colima (circles) and Popocatepetl (black diamonds). The regression lines are shown separately for large and small explosions. Dashed line with index N-H Law shows the scaling law proposed by Nishimura and Hamaguchi (1993). After Nishimura, T., Hamaguchi, H., 1993. Scaling law of volcanic explosion earthquakes. *Geophysical Research Letters* 20, 2479–2482 and Zobin, V.M., Reyes, G.A., Guevara, E., Bretón, M., 2009b. Scaling relationship for Vulcanian explosions derived from broadband seismic signals. *Journal of Geophysical Research* 114, B03203.

shows (Fig. 14.20) a strong dependence of the energy of the explosion on the counterforce (coefficient of correlation  $R=0.87$ ). In this case, both groups of large and small explosions demonstrate the same scaling of the magnitudes of the force controlling the magma movement before the explosion and the energy of the explosion.

Broadband seismic records of 14 explosions of Sakurajima volcano, discussed in Section 14.3.1, were used to obtain the relationship between the

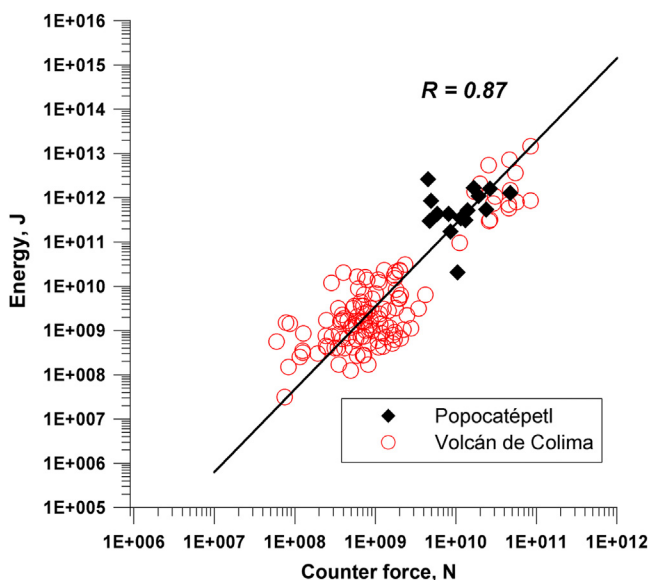


**FIGURE 14.20** Relationship between the counterforce of the eruption and the energy of the explosion for the Vulcanian explosions of Volcán de Colima. (circles) and Popocatepetl (black diamonds). From Zobin, V.M., Reyes, G.A., Guevara, E., Bretón, M., 2009b. Scaling relationship for Vulcanian explosions derived from broadband seismic signals. *Journal of Geophysical Research* 114, B03203.

seismic moment component  $M_{zz}$  of the preliminary LP phase estimated by Tameguri et al. (2002) and the energy of explosions estimated according to the methodology by Zobin et al. (2006a,b). The seismic records of station KUR situated at a distance of 4.1 km from the crater were used for estimation of the energy. Fig. 14.21A shows the typical seismic record of the Sakurajima explosion at station KUR and its main phase signal used for energy calculation. The relationship between  $M_{zz}$  and  $E$  is shown in Fig. 14.21B. The components  $M_{zz}$  of moment tensor, estimated by the inversion of the LP phase, are well correlated with the energy of small ( $E > 10^8$  J) explosions of Sakurajima. It allows obtaining the following equation relating these two parameters:

$$\log M_{zz} = 0.38 \log E + 8.4 \quad (14.5)$$

For microexplosions with energy  $E < 10^8$  J, the correlation between these two parameters was not observed.

#### 14.3.4.2 Strombolian Explosions

The relationship between the magnitudes of the initial and main stages of the Strombolian explosive process may be obtained, according to the two-stage



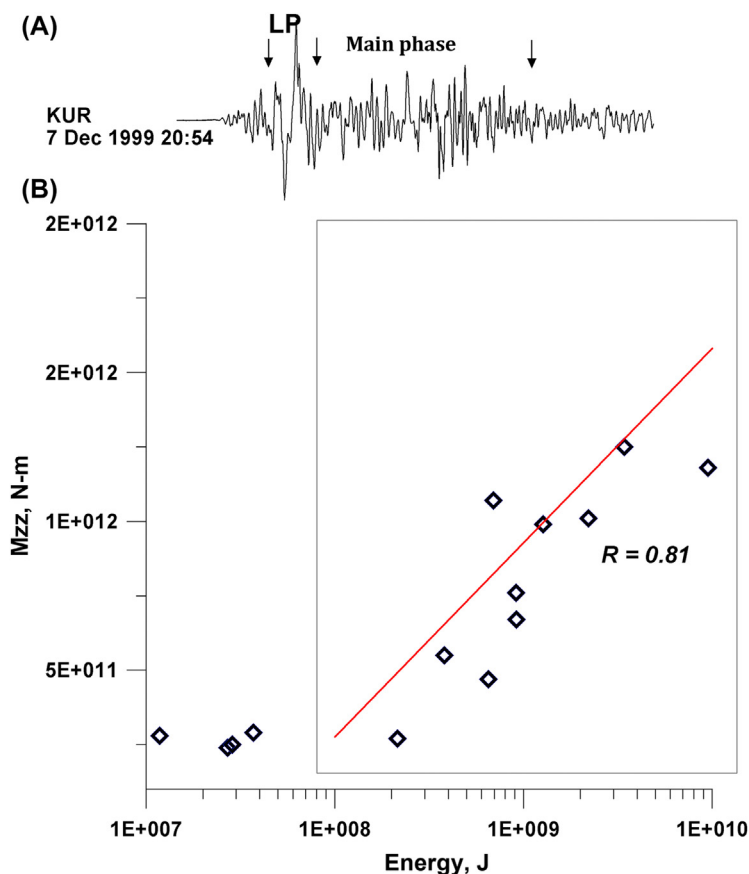


**FIGURE 14.21** Relationship between the components  $M_{zz}$  of moment tensors of Sakurajima explosions estimated in Tameguri et al. (2002) and the energy of the same events published in Zobin et al. (2013). The typical seismic signal used for estimation of energy is shown in (A); the plot of data is shown in (B). The small explosions are shown within the rectangle. These parameters are correlated with the coefficient of correlation  $R = 0.81$ . The correlation of these parameters for microexplosions ( $E < 108$  J) is absent. Courtesy: Sakurajima Volcano Research Center.

conceptual model (see Section 14.3.2), from the comparison between the seismic energies of the initial LP phase corresponding to the size of gas slug and the main seismic phase corresponding to the size of the explosion. Fig. 14.22 demonstrates the power-law distribution for the size parameters of the sources generating two phases of the seismic signals inferred from the parameters of 100 Strombolian explosions at the basaltic Aso volcano. This power-law distribution with the coefficient of correlation  $R = 0.82$  ( $R_{\text{crit}99\%} = 0.25$ ) indicates that the size of the gas slug strongly governs the energy of explosions.

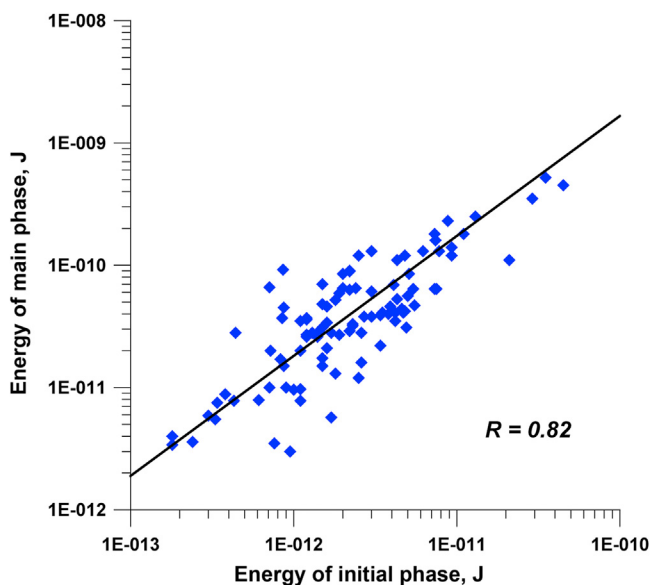


**FIGURE 14.22** Relationship between the seismic energy of two phases, initial and main, of the seismic signals of Strombolian events at Aso volcano. Short-period seismic station KAF, vertical component. The maximum likelihood regression line is shown.  $R$  is the coefficient of correlation. *Courtesy: Aso Volcano Observatory.*

## 14.4 LOCATION OF STROMBOLIAN AND VULCANIAN EXPLOSION EARTHQUAKES

As is seen from [Section 14.1](#), the seismic records of explosive events are rather complicated and their source may be considered as a sequence of subevents. Therefore it is important to locate these subevents, at least the position of the initial seismic source, preceding an explosion, and the position of the source of the explosion. The absence of good-expressed body waves in the seismic signal makes impossible the traditional hypocenter determination using onset arrival times.

### 14.4.1 Location of the Initial Subevents from Waveform Inversion

The waveform inversion, performed for the whole seismic record, gives the position of a centroid corresponding to the integrated source and cannot be applied here. Therefore only the results obtained for the initial phases of the seismic signal are discussed. [Kumagai et al. \(2011\)](#) performed the source location of the February 11, 2010, explosive event occurring at Tungurahua volcano, Ecuador, using waveform inversion for the initial part (about 5–7 seconds) of broadband seismic records. They obtained the

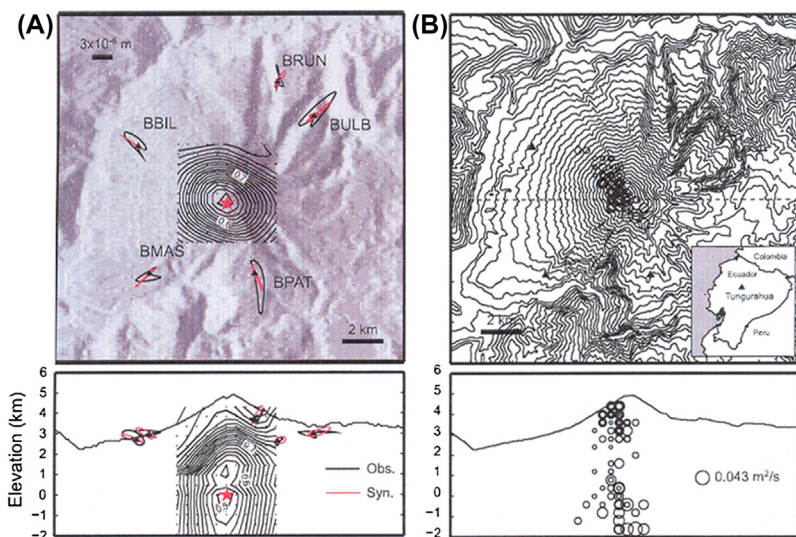


**FIGURE 14.23** Locations of broadband seismic and infrasonic stations. (triangles) at Tungurahua volcano, Ecuador, and contour plots of horizontal and vertical distributions of the normalized residuals obtained by waveform inversion of an explosion event on February 11, 2010. Observed (black lines) and synthetic (red lines, gray in print versions) particle motions obtained from displacement seismograms are also shown. Dots indicate the nodes used in the grid search to estimate the source location. (B) Source locations of the explosion event estimated by using vertical seismic amplitudes in a frequency band of 5–10 Hz. The circles showing source locations are scaled by the initial amplitudes. The horizontal dashed line B (top) is the horizontal location of the vertical profiles shown in A (bottom) and B (bottom). From Kumagai, H., Placios, P., Ruiz, M., Yepes, H., Kozono, T., 2011. Ascending seismic source during an explosive eruption at Tungurahua volcano, Ecuador. *Geophysical Research Letters* 38, L01306.

source position at a depth of about 5 km beneath the summit (Fig. 14.23) attributing it to the position of the initial ascending seismic source before the eruption.

## 14.5 THE SEQUENCES OF STROMBOLIAN AND VULCANIAN EXPLOSION EARTHQUAKES

The seismic records of explosion earthquakes form the families of similar waveforms (Johnson et al., 1998; Rowe et al., 2000). Strong repeatability of the gross waveform characteristics in the records of lava lake Strombolian explosions at Erebus volcano (Fig. 14.24) is evident among events up to 20 seconds into the coda, demonstrating a repetitive source process.

*Size–frequency relationship* for explosion earthquake sequences differs from the lineal regression common for volcano-tectonic earthquakes. Fig. 14.25 shows the size (maximum ground velocity) distribution of 364 explosions of Erebus volcano, Antarctica (Rowe et al., 2000). Note that the

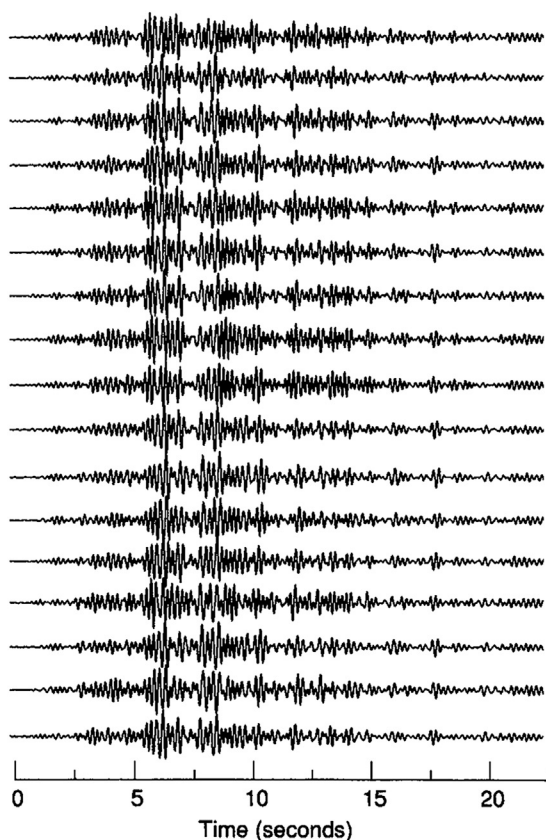


**FIGURE 14.24** Illustration of the similarity of seismic records for explosion earthquakes. Seventeen lava lake Strombolian explosions of Erebus volcano, Antarctica produced these short-period seismograms of explosion earthquakes that were recorded at a distance of 10 km from the lava lake. Traces have been bandpass filtered from 1 to 7 Hz to reduce incoherent signals. *From Rowe, C.A., Aster, R.C., Kyle, P.R., Dibble R.R., Schlue, J.W., 2000. Seismic and acoustic observations at Mount Erebus volcano, Ross Island, Antarctica, 1994–1998. Journal of Volcanology and Geothermal Research 101, 105–128.*

sharp bend in the cumulative curve of [Fig. 14.25B](#) corresponds to the peak in the histogram of [Fig. 14.25A](#). For 727 Karymsky explosion earthquakes, the magnitude (energy)–frequency relation had a parabolic form ([Fig. 14.26](#)). [Tokarev and Firstov \(1967\)](#) consider that the maximum ([Fig. 14.26](#)) corresponds to the optimal energy for explosions in the crater, according to the properties of the magmatic agents and conduit dimensions (“critical magma radius”; [Nishimura, 1995](#)); all remaining events represent the deviations from this optimal case.

The size–frequency distributions for large and small explosions, recorded at Volcán de Colima and defined in [Section 14.3.2](#), display the difference in

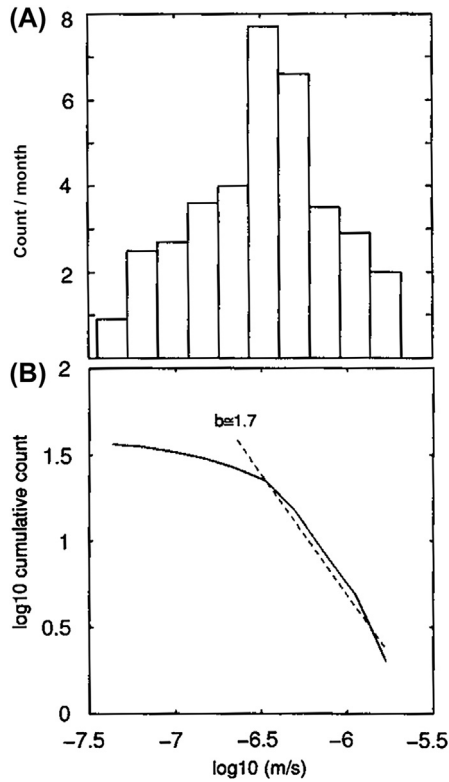


**FIGURE 14.25** Size (maximum ground velocity) distribution of 364 explosions of Erebus volcano, Antarctica recorded between October 1997 and July 1998 at a distance of 10 km from the lava lake. (A) Histogram of explosion counts versus  $\log_{10}$  of seismic amplitude, representative of explosion seismic magnitude. (B)  $\log_{10}$  of monthly cumulative counts (for each size bin and higher) versus  $\log_{10}$  seismic amplitude. From Rowe, C.A., Aster, R.C., Kyle, P.R., Dibble R.R., Schlue, J.W., 2000. Seismic and acoustic observations at Mount Erebus volcano, Ross Island, Antarctica, 1994–1998. *Journal of Volcanology and Geothermal Research* 101, 105–128.

their source properties. Fig. 14.27 shows the energy–frequency distributions obtained for two databases: (1) 15 large explosions ( $E > 10^{11}$  J) occurring during the March 2005 to September 2005 explosive sequence at Volcán de Colima and (2) 236 small explosions ( $E < 10^{11}$  J) of the November 2004 sequence at the same volcano. Application of the Kolmogorov–Smirnov test (Fig. 14.27B), considering as a null hypothesis that these two databases,  $n_1$  and  $n_2$ , belong to the same sample, demonstrated that the null hypothesis had to be rejected and these two databases belong to different samples. To reject the null hypothesis at the significance level  $\alpha$ ,  $D_{n_1, n_2}$  had to be greater than the critical value equal to  $k_1 - \alpha \sqrt{(n_1 + n_2)/n_1 n_2}$ , where  $k_1 - \alpha$  is the quantile of the Kolmogorov–Smirnov distribution at the significance level

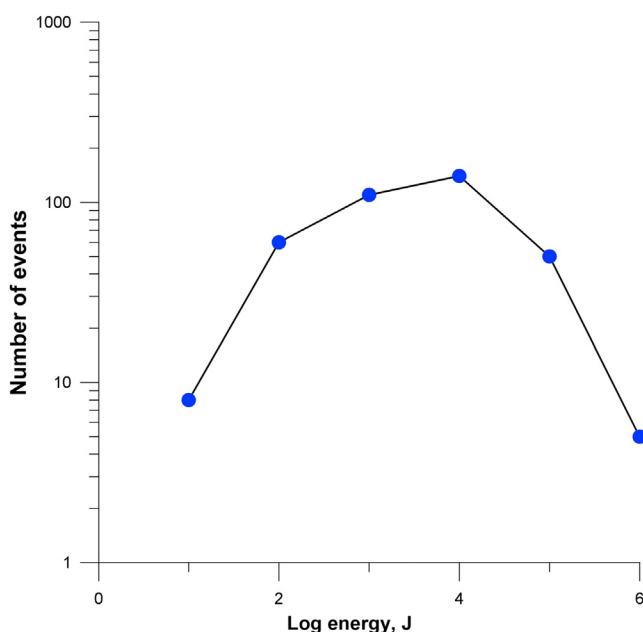


**FIGURE 14.26** Magnitude (energy in J). —frequency dependence for explosion earthquakes of Karymsky volcano. Data from Tokarev, P.I., Firstov, P.P., 1967. *Seismological study of Karymsky volcano. Bulletin of Volcanology Stanciy 43*, 9–22 (in Russian).

$\alpha$  (Muller et al., 1979). The maximum difference between the two cumulative distributions,  $D_{n1,n2}$ , was 1.00, which is larger than the 0.01—critical value of 0.4334. Therefore the large and small explosions at Volcán de Colima may be considered as having a different nature of generation.

## 14.6 STROMBOLIAN AND VULCANIAN EXPLOSION EARTHQUAKES IN ERUPTIVE PROCESS

Explosion earthquakes may be recorded at different stages of volcanic eruptions and give useful information about the process of magma emission.

### 14.6.1 Explosion Earthquakes as a Component of Eruptive Process

Fig. 14.28A shows the 10-year variations in the number of explosion earthquakes at Karymsky volcano. The mean rate of lava emission during each eruptive episode is shown in Fig. 14.28B. It is seen that the episodes with lava emission are characterized by an increase in the frequency of explosive earthquakes. At the same time, no dependence between the volume of lava emission and the maximum number of seismic events was observed (Khrenov et al., 1982).

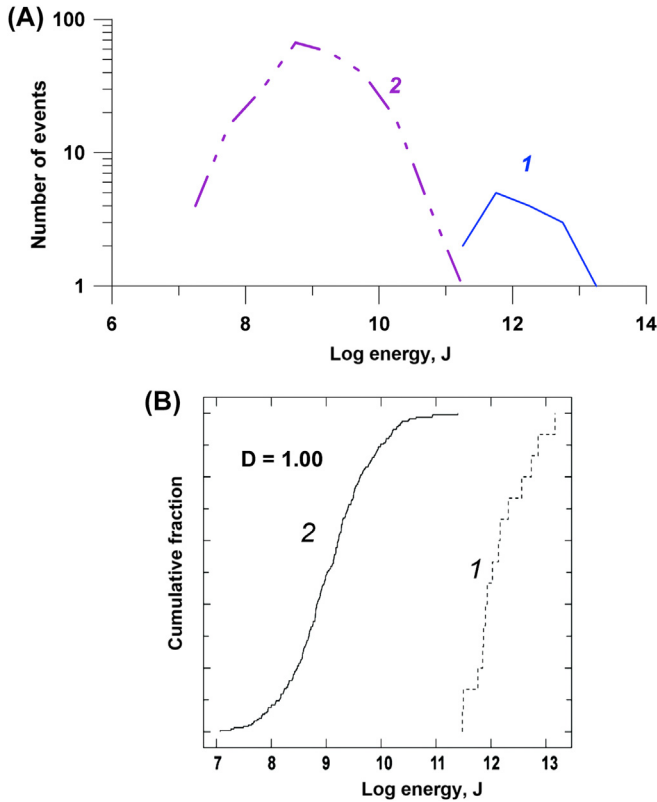


**FIGURE 14.27** Distribution in the energy for large explosions of the March 2005 to September 2005 sequence (1) and small explosions of the November 2004 sequence (2) at Volcán de Colima (A) and the Kolmogorov–Smirnov test comparison cumulative fraction plot for the sequences 1 and 2 (B). The statistic  $D$  is the maximum vertical deviation between two curves. *Courtesy: Colima Volcano Observatory.*

Tanaka et al. (1972) studied the explosion earthquakes that occurred during the 1970–71 eruption of the Akita-Komaga-take volcano, in Japan. The eruption began with a sequence of Strombolian explosions and basaltic lava emission on September 18, 1970, and continued until January 25, 1971. About 33,000 explosion earthquakes were recorded. They were the main seismic signals for this eruption. Single volcano-tectonic earthquakes (B) were very innumerable showing the variation in the number of explosion earthquakes that were observed from the very beginning of the eruption. These earthquakes were directly connected with volcanic explosive activity and may serve for monitoring of this type of eruption Fig. 14.29.

Fig. 14.30 shows the effusive–explosive activity of Volcán de Colima during 2002–06. During this period, two episodes of lava flow activity were recorded, from February 2002 to February 2003 and from September 2004 to

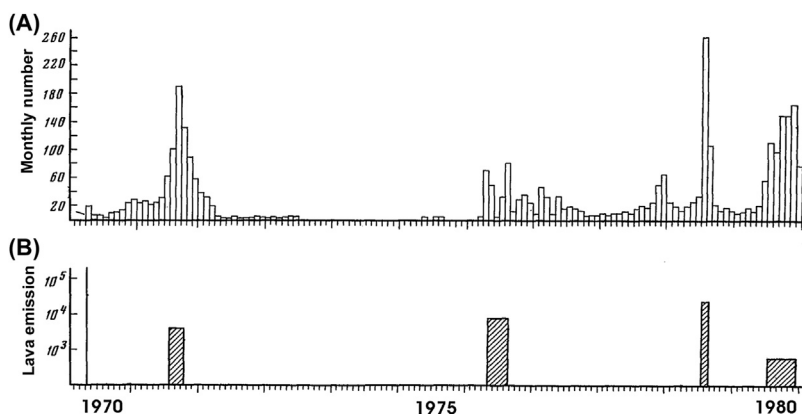


**FIGURE 14.28** Variations in the monthly number of explosion earthquakes (A) and in the mean rate of lava emission (in kg/s) during eruptive episodes of Karymsky volcano, Kamchatka, during 1970–80 (B). From Khrenov, A.P., Dubik YuM., Ivanov, B.V., Ovsyannikov, A.A., Pilipenko, V.P., Taran Yu A., Firstov P.P., Chirkov, A.M., 1982. Eruptive activity of Karymsky volcano in 1970–1980. *Journal of Volcanology and Seismology* 4, 29–48 (in Russian).

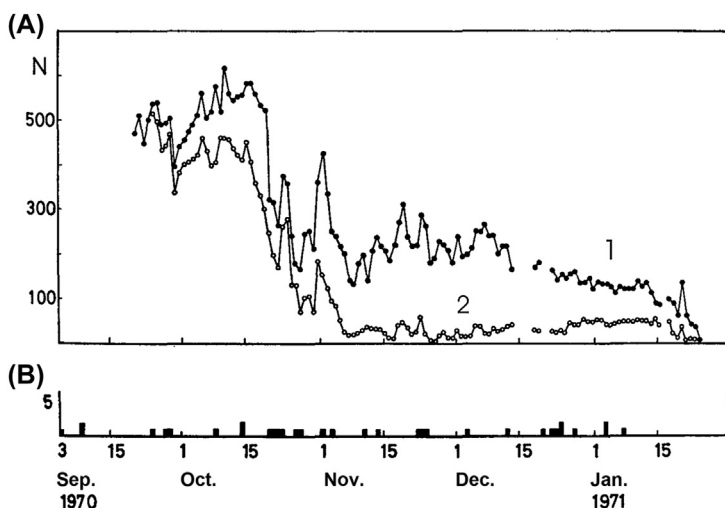


**FIGURE 14.29** Daily number of explosion earthquakes (A) and volcano-tectonic earthquakes with S-P times less than 3.0 s (B) recorded during the eruption of Akita-Komaga-take volcano, Japan, at the seismic station situated at a distance of about 4 km from the crater. Open circles (curve 2) represent the number of explosion earthquakes with maximum amplitude larger than  $200 \mu\text{m}$ . From Tanaka, K., Kasahara, M., Hori, S., Suzuki Z., and Takagi, A. (1972). *Research on Akita-Komaga-take (I). Summary of its eruptions in 1970–1971. Sci. Rep. Tohoku Univ., Ser. 5* 21, 61–75.



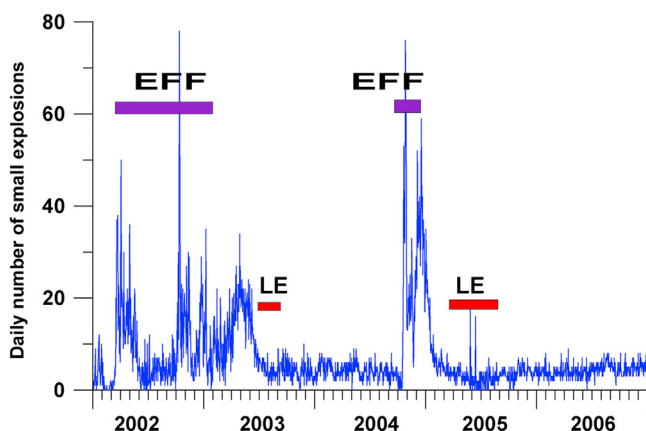


**FIGURE 14.30** Development of effusive–explosive activity of Volcán de Colima during 2002–06. Variations in the daily number of small explosions are shown together with the intervals of lava extrusions (EFF) and large Vulcanian explosions (LE). *Courtesy: Colima Volcano Observatory.*

December 2004. The increase in the number of small explosions, recorded at a distance of 1.7 km from the crater by a short-period seismic station, was observed during these two episodes. Large explosions (energy greater than  $10^{11}$  J) occurred when the effusive activity was terminated and the activity of small explosions significantly decreased.

Zobin et al. (2006a,b) showed that the explosions at Volcán de Colima were accompanied by pyroclastic flows only if their energy was greater than  $10^{11}$  J. It is illustrated in Fig. 14.31.

### 14.6.2 Vulcanian Explosions as Indicators of the Style of Eruption Activity

The difference in the nature of the generation of three types of Vulcanian explosions, described in Section 14.1.2, may be used for the search of indicators of different styles of eruption during the process of building or destruction of the lava dome at an andesitic volcano. It was illustrated by Zobin et al. (2015) for two episodes of the 1998–2011 eruption at Volcán de Colima. The first was the episode of posteffusive explosions recorded during March 2003 to August 2003 when a 6-month stage of small explosions followed the ceasing of lava dome building in the crater. The lava dome was destroyed by large August 2003 explosions. The second episode, selected for analysis, was the episode of coeffusive explosions recorded from January 2007 to June 2007 at the initial 6-month stage of the 2007–11 slow growth

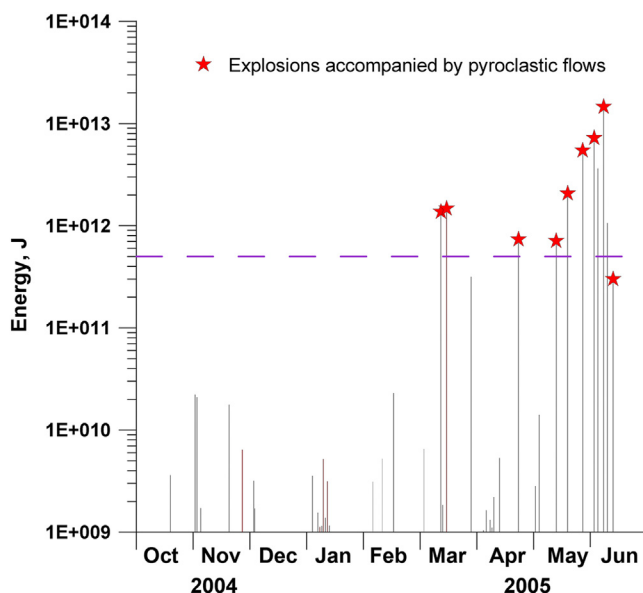


**FIGURE 14.31** The energy of Vulcanian explosions observed at Volcán de Colima during 2004–05 (shown as bars), and the appearance of pyroclastic flows accompanying the largest of them (with energy  $>10^{11}$  J). *Courtesy: Colima Volcano Observatory.*

of a new lava dome in the crater. The comparative study of temporal variations in the number of three types of explosive events demonstrates the difference in the nature of explosive activity during these two stages.

#### 14.6.2.1 Posteffusive Explosions

Fig. 14.32A shows the temporal variations of explosive events of three types from March 2003 to August 2003. The explosive events were distributed unevenly in time, changing from 22 to 59 monthly events. From a total of 257 identified events, 152 (59%) were of Type 1, 12 (5%) events of Type 2, and 93 (36%) events of Type 3. Type 1 and Type 3 signals represented the majority (95%) of explosive waveforms. The number of Type 1 events gradually increased from March to July but then decreased before the appearance of large August explosions. Type 3 events sharply decreased in number earlier, beginning in June. The variation in the low-number Type 2 explosive waveforms, supposedly generated within the lava dome body, was not significant.

The predominance of Type 1 and Type 3 signals and their temporal distribution may indicate, according to the conceptual model (Section 14.3.2), that the posteffusive stage was characterized by the intensification of processes within the conduit before large explosions.

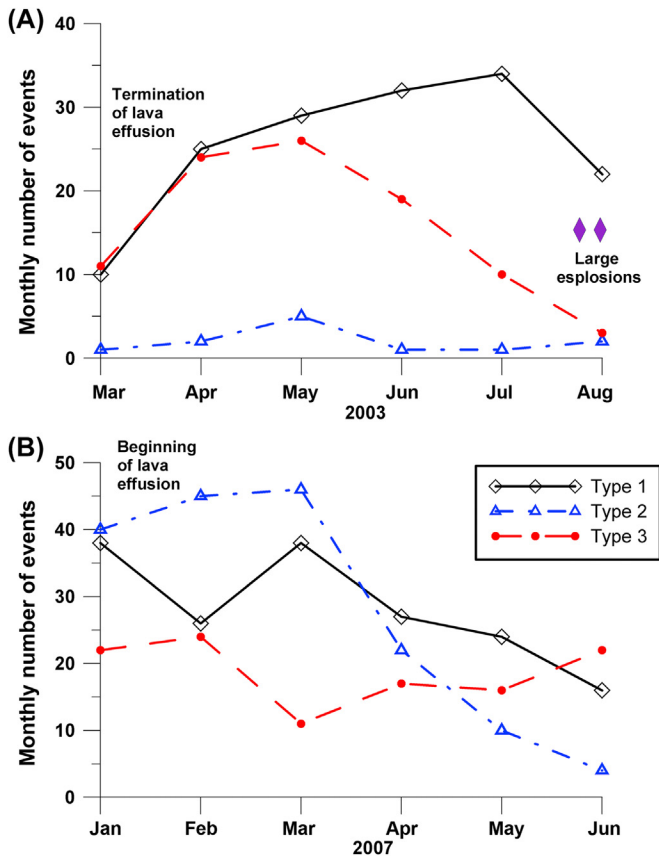


**FIGURE 14.32** Two 6-month stages of the eruption at Volcán de Colima selected for analysis: (A) the 2003 posteffusive stage culminated with the two large (with energy  $\geq 1012$  J) August explosions and (B) the 2007 coeffusive stage during the beginning of new lava dome growth in the crater of the volcano. In (A), the end of lava effusion is indicated with the termination of rockfalls associated with the lava dome growth. 1, daily frequency of rockfalls; 2, daily frequency of small explosions. The moments of large explosions are shown with diamonds. In (B), the selected stage of the beginning of slow lava dome growth is shown with the parallelogram. The volume of the growing lava dome was estimated from the sequences of aerial photos and subsequent field observations of the lava flow contours and thickness (BGVN, 2008). From Zobin, V.M., Peral, J.J., Nava, F., Bretón, M., 2015. Vulcanian explosions in the process of building-destruction of the lava dome of andesitic volcano: insight from the seismic signals recorded at Volcán de Colima, México. *Physics of the Earth and Planetary Interiors* 244, 1–10.

### 14.6.2.2 Coeffusive Explosions

A total of 448 seismic records were identified as produced by explosive events during the first 6 months of the initial stage of the 2007 lava extrusion from January 2007 to June 2007. They were distributed slowly decreasing in time, from 95 to 100 identified events in January–March to 42–66 monthly

events in April–June (Fig. 14.32B). Of these 448 events, 169 (38%) were of Type 1, 167 (37%) events of Type 2, and 112 (25%) events of Type 3. The majority of explosive events were Type 1 and Type 3 events (63%) but the number of Type 2 events became significant (37%) when compared to that in the posteffusive stage. The beginning of a new lava dome growth in January–March was marked by a predominance of Type 2 explosive events; then their number began to gradually decrease. The number of Type 1 explosions was also rather high during January and March. The number of Type 3 events was stably low during all 6 months of observations.

The predominance of the Type 2 explosive waveforms during January–March, supposedly generated during the fracturing of the subsurface lava cap, may serve to open a way for the magma ascent to the surface and initial lava dome building.

## 14.7 NATURE OF SEISMIC SIGNALS OF EXPLOSION EARTHQUAKES OCCURRING DURING PLINIAN ERUPTION AND THEIR CLASSIFICATION

In this section, I discuss the nature of the seismic events recorded during the Plinian eruption at the HTHH submarine volcano on January 15, 2022. The process of development of the Plinian explosive event, derived from the seismic data, and quantification of the overall mechanical impulse, generated by the explosion, are presented.

### 14.7.1 Process of Development of the Plinian Explosive Event Derived from the Seismic Data

As was noted in Section 14.1.2 and shown in Fig. 14.6, four remarkably similar seismic subevents, E-1 to E-4, occurred within a 5-minute interval during the intensifying eruptive phase and were recorded by global seismic stations (Matoza et al., 2022; Yuen et al., 2022; Zheng et al., 2023). Detailed seismic analyses showed that each of these subevents was similar in waveform and duration ( $\sim 25$  seconds) and was characterized by a sequence of four forces: upward, downward, upward, and downward. Considering that the four seismic subevents exhibited similar waveforms, Zheng et al. (2023) realized the detailed analysis for only the E-1 subevent time series, from stages A to D as shown in Fig. 14.33. The respective durations for stages A to D (Fig. 14.33B) were estimated as  $\sim 3$  seconds, 15 seconds, 3 seconds, and 3 seconds. The magma hammer was selected as one possible mechanism to cause the single forces. The following modeling of the four stages was proposed.

*Stage A* lasted about  $\sim 3.2$  seconds and the far field ground motion was characterized by a distinct downward ground motion, meaning that the seismic source was likely to be an upward single force. Full-wave seismic modeling revealed that this upward single-force source had a force

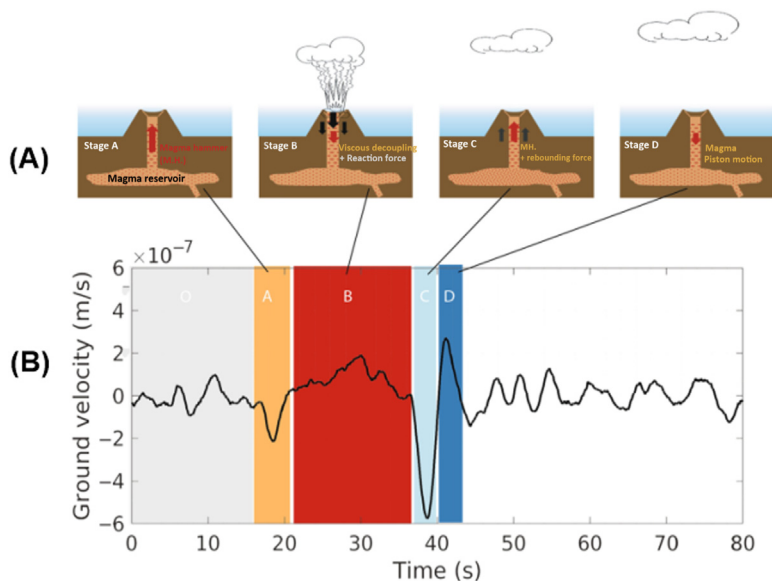


**FIGURE 14.33** Eruption stages and magma plumbing system of the HTHH eruption. (A) Volcano-reservoir schematics of states and forces (red arrow: magma hammer; black arrow: reaction force) corresponding to Stages A–D. The magma reservoir is centered at a depth of 5–8 km based on geobarometry. (B) Globally stacked ground velocity (vertical component) of event E-1. The amplitude of the waveform is not linearly proportional to the force history. Time zero corresponds to ~04:14:45 January 15, 2022. From Zheng, Y., Hu, H., Spera, F. J., Scruggs, M., Thompson, G., Jin, Y., et al. (2023). Episodic magma hammers for the 15 January 2022 cataclysmic eruption of Hunga Tonga-Hunga Ha’apai. *Geophysical Research Letters*, 50, e2023GL102763. <https://doi.org/10.1029/2023GL102763>.

magnitude  $\sim 5.0 \times 10^{12} \text{ N}$ . It was proposed that Stage A represented the magma hammer breaching of a short-lived flow blockage of the magma conduit such as a caldera plug, generating an upward single force.

Stage B was the longest stage (15 seconds) (Fig. 14.33A). It is likely due to the reactive impulsive force associated with the opening of the magmatic conduit and transient reestablishment of intense outflow for  $\sim 15$  seconds (Fig. 14.33A). During this stage, the eruption could have sent a series of pulses of volcanic product into the atmosphere based on observed maximum plume heights. This “upside-down rocket”-type eruption created a downward single force on the order of  $\sim 9 \times 10^{12} \text{ N}$ . In Stage B, the net force first increases and then decreases, and both upward force and downward forces may coexist. Therefore the downward forces dominated toward the early part of the stage.

Stages C and D, initiated about 20 seconds after the onset of Stage A, exhibited an upward force in C of  $\sim 4 \times 10^{12} \text{ N}$ , and a downward force in D, equal to  $\sim 3 \times 10^{12} \text{ N}$ . Stage C was associated with a second-phase

closing/constricting of the conduit and the subsequent generation of an upward magma-hammer force or viscous drag upward force. It is likely that Stage C was related to the magma backflow impacting the overburdened rock containing the constricted conduits, which creates an upward magma-hammer force and an upward viscous drag force due to the flow velocity. Stage D seems to be a reflected magma-hammer signal traveling down, which creates a downward force as viscous drag or impact of the magma chamber. Stages C and D represented one cycle of oscillatory piston motion as magma motion travels through the conduit, from the reservoir to the surface.

### 14.7.2 Quantification of the Plinian Explosive Process

The quantification of the explosive Plinian process was based on back projection of long-period  $P$  and surface Rayleigh waves, recorded by the global seismic network (Poli and Shapiro, 2022). They proposed an approximation of the eruptive process as a vertical explosion and force, which was plausible at first order considering the nearly radially symmetric shape of the HTHH ash cloud seen by satellites (see Fig. 14.6A).

The spectral representation of the source force  $F_z(\omega)$  was computed with a Fourier transform:

$$F_z(\omega) = \int f_z(t) e^{-i\omega t} dt \quad (14.6)$$

If the explosion time history  $F_z(t)$  is positively defined, its Fourier spectrum  $F_z(\omega)$  will converge at low frequencies to a “plateau”:

$$\lim_{\omega \rightarrow 0} F_z(\omega) = \int F_z(t) dt = K_0 \quad (14.7)$$

where  $K_0$  is the mechanical impulse of the material ejected by the explosion (volcanic explosion impulse). If the force can be approximated with a single unidirectional pulse of duration  $\tau_{\text{exp}}$ , the spectral amplitude falls down above the corner frequency  $f_c$  approximately inverse to the explosion duration:

$$f_c \approx 1/\tau_{\text{exp}} \quad (14.8)$$

The obtained force spectrum of Rayleigh waves, representing the average of the 41 single stations, is shown with a blue line in Fig. 14.34. It was obtained after the deconvolution of 6000 seconds of signals. The force spectrum shows a low-frequency spectral “plateau” for frequencies lower than 5 mHz. The level of this plateau was used to estimate the overall mechanical impulse of the explosion  $K_0$ , equal to  $\approx 1.3 \times 10^{15}$  Ns. This value is  $\sim 2.5$  times larger than the estimate of Kanamori and Given (1982) for the 1980 Mount St Helens eruption (see green line in Fig. 14.34).

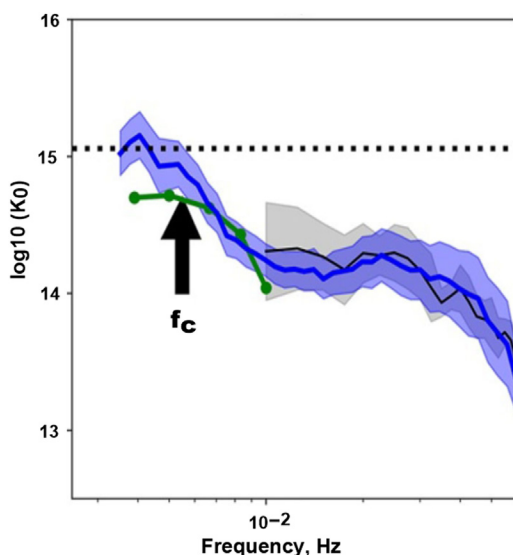


**FIGURE 14.34** Source spectrum, obtained from the deconvolution of 6000 s of signal, containing Rayleigh waves (*continuous blue line*) and its confidence bound (*blue area*). The *black line* and *gray area* show the source spectrum obtained using P-waves. The arrow indicates the estimated corner frequency  $f_c$ . The *dashed line* shows the level of estimated volcanic explosion impulse  $K_0 = 1.3 \times 10^{15}$  N s. The force spectrum for the 1980 eruption at Mt. St. Helens (Kanamori and Given, 1982) is shown with a *green line*. From Poli, P., Shapiro, N.M., 2022. Rapid characterization of large volcanic eruptions: measuring the impulse of the Hunga Tonga Ha’apai explosion from teleseismic waves. *Geophys. Res. Lett.* 49. e2022GL098123. <https://doi.org/10.1029/2022gl098123>.

The spectrum started to fall off at a corner frequency  $f_c \approx 0.005$  Hz, suggesting an approximate explosion duration of  $\sim 200$  seconds Eq. (14.8).

The average force spectrum from P-waves (black line in Fig. 14.34) is remarkably similar to the one from Rayleigh waves at frequencies above 0.01 Hz. At the same time, it is seen that the spectral falloff above the corner frequency is not “homogeneous” with the appearance of a second plateau between 0.01 and 0.04 Hz. This specific shape of the spectrum of P-waves allows additional estimation of  $K_0 \approx 2 \times 10^{14}$  Ns and a corner frequency close to 0.04 Hz.

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## Chapter 15

# Long-Period and Very-Long-Period Seismic Signals at Volcanoes

This chapter is devoted to the seismic signals that are recorded at volcanoes without any association with a particular type of eruptive event but they are common for volcanic activity in general. They are the so-called long-period and very-long-period seismic signals. The seismic signals recorded at volcanoes with an emergent onset of P-waves, no distinct S-waves, and a harmonic signature in the range of periods between 0.2 and 2 seconds (5 and 0.5 Hz) are defined by Chouet (2009) as long-period (LP) seismic signals. They are characterized by simple decaying harmonic oscillations except for a brief time interval at the event onset (Chouet, 2003). Seismic signals in the range of periods between 2 and 100 seconds (0.5 and 0.01 Hz) are defined as very-long-period (VLP) seismic signals (Chouet, 2009). These signals became a subject of intense study with the beginning of broadband seismic observations at volcanoes in the 1990s. They were recognized at different volcanoes and during different stages of eruptions.

## 15.1 WAVEFORMS AND SPECTRA

### 15.1.1 Long-Period Seismic Signals

LP seismic signals, derived from volcanic seismicity and used by scientists, may represent the natural LP seismic records or LP component of raw signals of volcanic earthquakes obtained after filtering. Fig. 15.1 shows the typical waveforms of the natural LP seismic signals and their Fourier spectra.

The seismic signals, demonstrated in Fig. 15.1, were recorded during different stages of eruptive activity. The first of them (Fig. 15.1A) was observed simultaneously with a swarm of volcano-tectonic earthquakes preceding the 2000 flank eruption at Usu volcano (see Section 7.2). It has an emergent high-frequency onset; the body waves are not identified. The amplitude of the signal slowly increases and then gradually decreases. These LP signals began to appear within the swarm when the deformation of

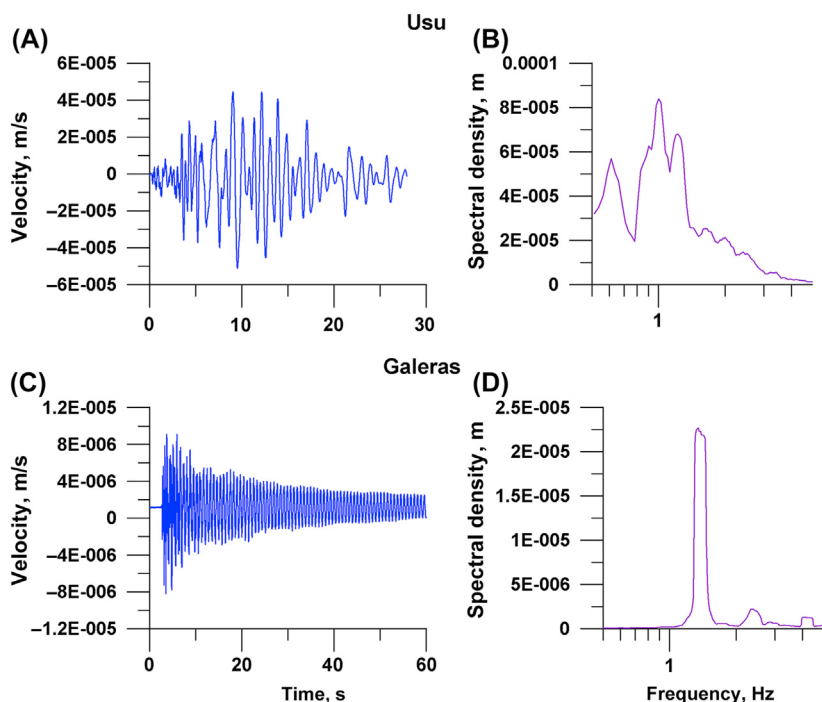


**FIGURE 15.1** Examples of long-period seismic waveforms (velocity, vertical component) and their Fourier spectra. (A) Seismic signal recorded at Usu volcano on March 30, 2000, and its Fourier spectrum (B). (C) Seismic signal recorded at Galeras volcano on January 1, 2010, and its Fourier spectrum (D). *Courtesy: (A) Japan Meteorological Agency; (C) INGEOMINAS, Colombia.*

the site of the forthcoming flank eruption became significant indicating a magma intrusion (Zobin et al., 2005).

The second signal is the so-called “tornillo” (screw; Fig. 15.1C); it was recorded on January 1, 2010, before the January 2 explosive eruption at Galeras volcano. Signals of this type show emergent onsets but occasionally may have slight impulsive onsets. They are characterized by slowly decaying coda waves. These signals are the characteristics of different volcanoes of the world and may be recorded before the eruptions, during the swarms of volcano-tectonic earthquakes, after eruptions, or during quiescent periods (Gómez and Torres, 1997).

All these events are characterized by the peak spectra between 0.8 and 2 Hz (Fig. 15.1B and D).

### 15.1.2 Very-Long-Period Seismic Signals

VLP seismic signals are characterized by impulsive signatures; their common characteristic is a period larger than 2 seconds (Fig. 15.2, Fig. 15.3, Fig. 15.4).

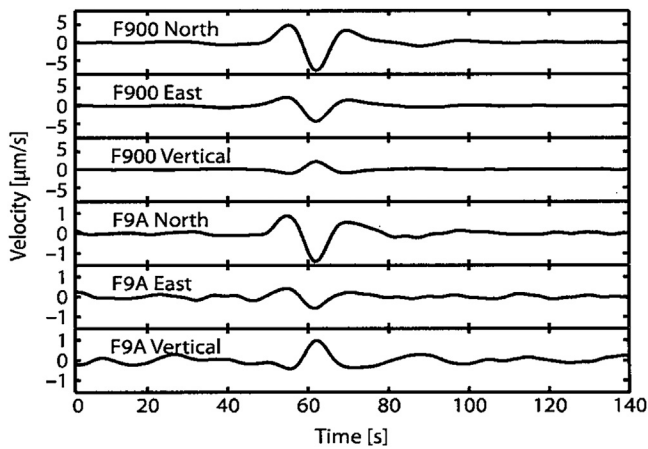


**FIGURE 15.2** VLP seismograms of the January 2009 explosion at Fuego volcano, Guatemala, extracted from the Guralp CMG 40 T (0.02–30 s) broadband seismic records. The seismic stations were situated at distances of 800 m (F900) and 1.2 km (F9A) from the crater. *After Lyons, J.J., Waite, G.P., 2011. Dynamics of explosive volcanism at Fuego volcano imaged with very long period seismicity. Journal of Geophysical Research 116, B09303.*

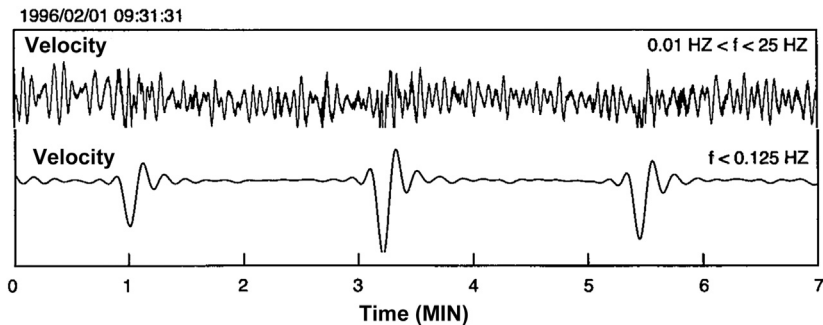


**FIGURE 15.3** Broadband velocity record. (*top trace*) and associated filtered signal of very-long-period (VLP) events (*bottom trace*) recorded at Kilauea volcano during a surge in the magma flow feeding the east rift eruption on February 1, 1996. *After Dawson, P.B., Dietel, C., Chouet, B.A., Honma, K., Ohminato, T., Okubo, P., 1998. A Digitally Telemetered Broadband Seismic Network at Kilauea Volcano, Hawaii. U.S. Geological Survey. Open File Report 98–108, 1–121.*

VLP signal may be seen frequently only after filtering of the raw signal. Fig. 15.2 shows the VLP signals obtained during the January 2009 explosions at Fuego volcano, Guatemala. These signals were extracted from the Guralp CMG 40 T sensors (0.02–30 seconds) records by deconvolving the instrument response and then filtering with a two-pole, zero-phase-shift Butterworth band-pass filter with corners at 30 and 10 seconds. The highest energy portion of the VLP signal lasts at least 30 seconds (Lyons and Waite, 2011). Fig. 15.3 shows the repetitive VLP signals recorded at the Kilauea volcano (the fourth

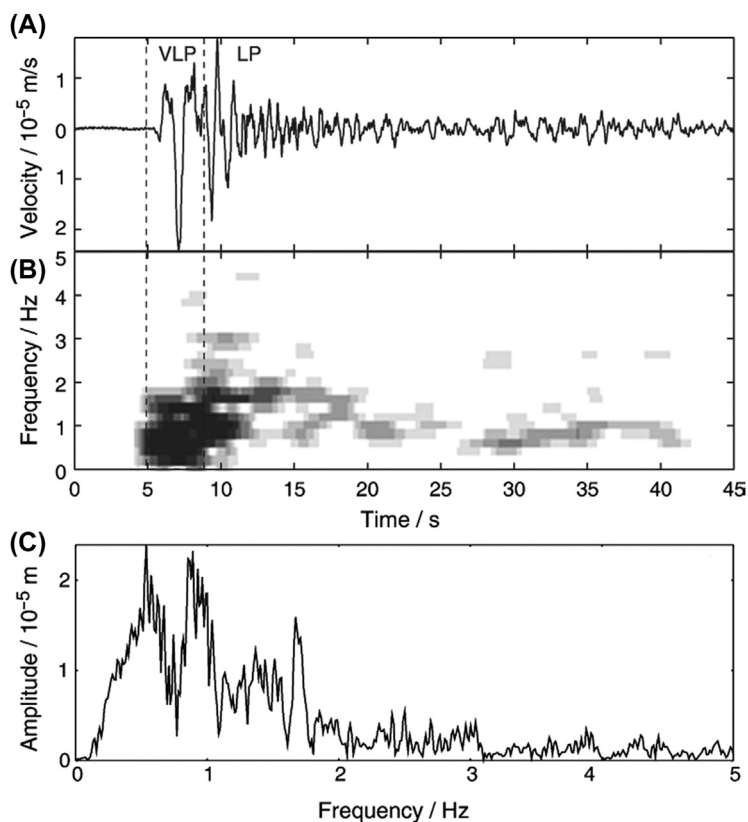


**FIGURE 15.4** Broadband velocity very-long-period (VLP) seismic record accompanied by long-period (LP) signal (A) recorded at Cotopaxi volcano on June 26, 2002, without any reported surface manifestation of volcanic activity, its spectrogram (B) and Fourier spectrum (C). From Molina, I., Kumagai, H., García-Aristizábal, A., Nakano, M., Mothes, P., 2008. Source process of very-long-period events accompanying long-period signals at Cotopaxi volcano, Ecuador. *Journal of Volcanology and Geothermal Research* 176, 119–133.

trace). These signals, consisting of pulses with periods of about 20 seconds, were recognized only when a low-pass filter ( $f < 0.125$  Hz) was applied. VLP signals may be followed by LP signals (Fig. 15.4).

VLP signals, shown in Fig. 15.2, Fig. 15.3, and Fig. 15.4, were recorded during different stages of eruptive activity such as explosive activity at Fuego volcano (Fig. 15.2), a surge in the magma flow at Kilauea volcano (Fig. 15.3), or without any superficial activity at Cotopaxi volcano (Fig. 15.4). Fig. 15.5 illustrates the dependence of the amplitude of VLP signals on the intensity of the explosive process that was observed at Santiaguito volcano, Guatemala (Sanderson et al., 2010). Two volcanic events, a small degassing and a Vulcanian explosion, occurred on January 1, 2009 (Fig. 15.5A). The seismic signals, associated with these two events, were recorded by broadband seismic

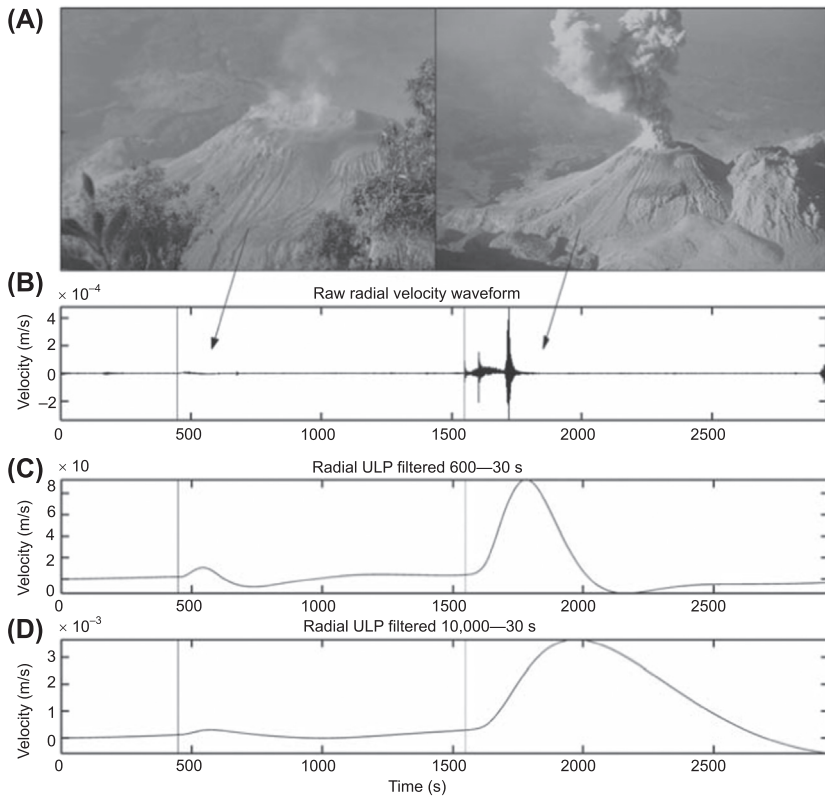


**FIGURE 15.5** Dependence of the very-long-period (VLP) or ultra-long-period (ULP) seismic signal amplitude on the intensity of explosive process at Santiaguito volcano. (A) Photos of small degassing (left) and Vulcanian explosion (right) occurring on January 1, 2009. (B) Raw seismic broadband record. (C, D) Bandpass filtered VLP seismic signals. Vertical bars mark the onset of eruption manifestations for each of the event. After Sanderson, R.W., Johnson, J.B., Lees, J.M., 2010. Ultra-long period seismic signals and cyclic deflation coincident with eruptions at Santiaguito volcano, Guatemala. *Journal of Volcanology and Geothermal Research* 198, 35–44.

station at a distance of about 500 m from the crater (Fig. 15.5B). These signals were bandpass filtered between 600 and 30 seconds (Fig. 15.5C) and between 10,000 and 30 seconds (Fig. 15.5D). The filtered seismic records displayed VLP signals associated with two volcanic events.

### 15.1.3 Occurrences of Long-Period and Very-Long-Period Events

LP and VLP events often occur in sequences where the events are very similar and the time separating events remains relatively constant. Fig. 15.6 shows the occurrence of LP (1–2 Hz) events observed at Shishaldin volcano, Aleutian Islands (Petersen et al., 2006). These events began to occur at the

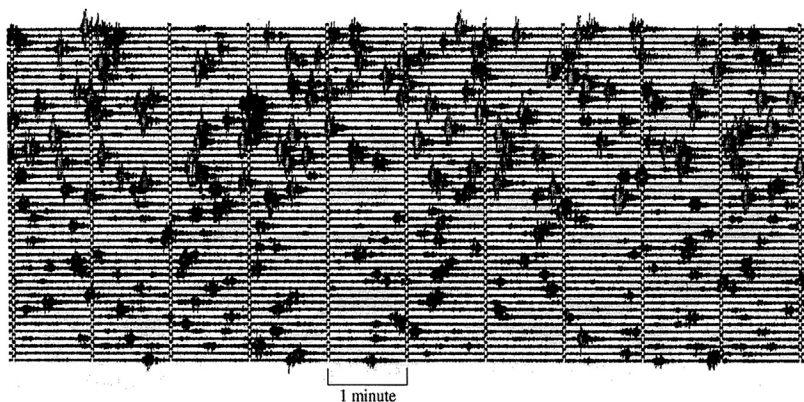


**FIGURE 15.6** Occurrence of long-period (LP) events recorded at Shishaldin volcano, Aleutian Islands, on November 21, 2002. The section of the short-period helicorder record shows 8 h of data recorded at a distance of 6.3 km from the summit. *From Petersen, T., Caplan-Auerbach, J., McNutt, S.R., 2006. Sustained long-period seismicity at Shishaldin volcano, Alaska. Journal of Volcanology and Geothermal Research 151, 365–381.*

volcano from September 1999, a few months after the April 1999 to May 1999 eruption. They continued until April 2004. This activity consisted of many hundreds to thousands of events per day, occurring every 1–2 minutes. The LP waveforms formed families consisting of nearly identical signals for time spans of days to months (Fig. 15.7A).

Long-term sequences of the LP events with similar waveforms have been observed at many volcanoes. At Mount Pinatubo, there were three periods with the occurrences of similar waveforms: 1 week before the June 1991 cataclysmic eruption, a day before a new 1992 dome began to grow, and during the 1992 dome growth (Ramos et al., 1999).

Numerous LP signals (Fig. 15.7B) were recorded during the 1989–90 Redoubt, Alaska eruption (Chouet et al., 1994). The initial eruption on December 14, 1989, was heralded by a 23-hour swarm of LP events that ended abruptly with the eruption. Another swarm, with LP events similar to those of the first, began on December 26 and ended with a major tephra eruption on January 2, 1990. A significant but less intense swarm preceded the February 15, 1990 eruption. From March 4 to the end eruption on April 21, six swarms of LP were recorded. All of them preceded the eruption episodes.

LP events accompanied the eruptive activity at Popocatepetl volcano from December 1994 through May 1995 (Arciniega-Ceballos et al., 2003); they were recorded in 1991 during the emplacement of a lava dome in Galeras volcano (Gil Cruz and Chouet, 1997), and they were reported during the 2004 observations at Etna volcano (Lokmer et al., 2007). The number of references may be continued.

Families of VLP signals have been recorded at different volcanoes: Stromboli (Chouet et al., 2003), Popocatepetl (Arciniega-Ceballos et al., 1999),



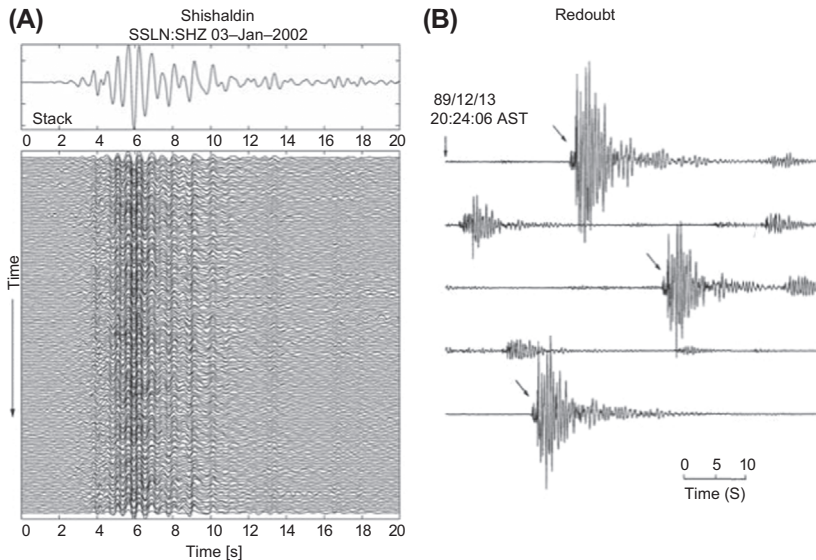


**FIGURE 15.7** Examples of repeating events at Shishaldin volcano recorded at a distance of 6.3 km from the summit (January 3, 2002) (A) and at Redoubt volcano recorded at a distance of 4 km from the summit (B). The cluster at Shishaldin volcano was extracted by cross-correlation. The spectral-coherence values for these events were  $\geq 0.9$ . The upper part of the plot in (A) shows a stack of all events in the lower part. The similar waveforms in (B) are indicated by arrows. Data from Chouet, B., Page, R., Stevens, R., Lahr, J., Power, J., 1994. Precursory swarms of long-period events at Redoubt volcano (1989–1990), Alaska: their origin and use as a forecasting tool. *Journal of Volcanology and Geothermal Research* 62, 95–135.

and Erebus (Aster et al., 2008), among others. Fig. 15.8 shows three families of VLP seismic records from Strombolian eruptions of the Erebus volcano. Group 1 eruptions (Fig. 15.8A) are characterized by positive initial vertical motions, and group 2 eruptions (Fig. 15.8B) show negative initial vertical motions. Group 3 eruptions (Fig. 15.8C) were very rare and exhibited a relatively simple pulse-like shape.

#### 15.1.4 Nature of Long-Period and Very-Long-Period Seismic Signals

The main properties of LP and VLP signals are summarized as follows: they are repetitive; they form families consisting of similar waveforms, which can be sustained for a long time. The presence of long-lived families of similar waveforms demonstrates the stable and nondestructive nature of their sources. Chouet (2003) noted that the LP waveforms, characterized by simple decaying harmonic oscillations, may be interpreted as oscillations of a fluid-filled oscillator in response to a time-localized excitation and can be thought of as the manifestation of acoustic resonance. Unlike LP waveforms, VLP signals may be thought of as the result of inertial forces

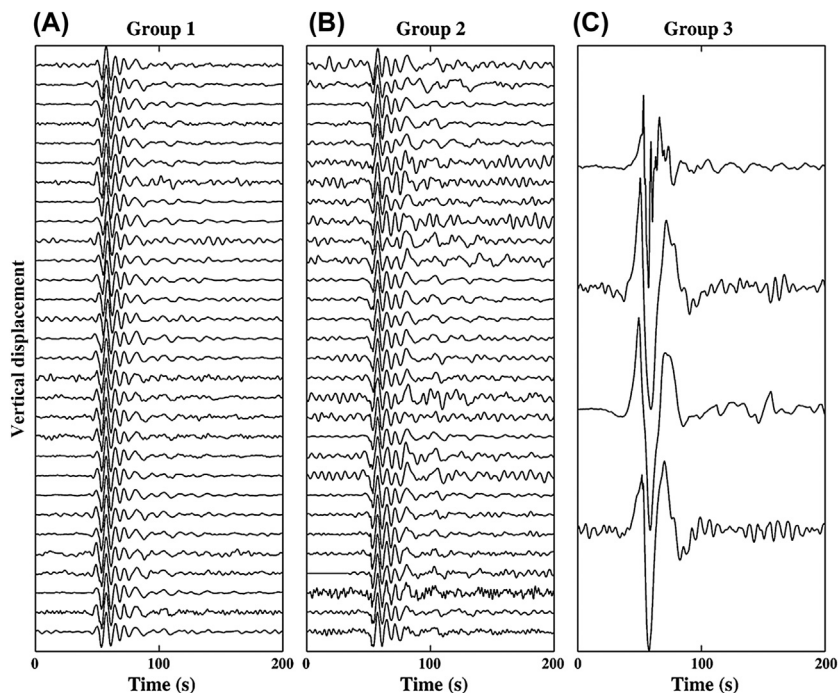


**FIGURE 15.8** Three families of very-long-period (VLP) seismic signals recorded during Strombolian explosions at the lava lake of Erebus volcano. From Aster, R., Zandomenighi, D., Mah, S., McNamara, S., Henderson, D.B., Knox, H., et al., 2008. Moment tensor inversion of very long period seismic signals from Strombolian eruptions of Erebus Volcano. *Journal of Volcanology and Geothermal Research* 177, 635–647.

associated with perturbations in the flow of magma and gases through conduits (Chouet, 2009). Both these types of sources may be considered as multiphase systems.

#### 15.1.4.1 Seismic Waves Generated within the Two-Phase System

The fluid-filled oscillator may be represented as a solid–fluid composite (Kawakatsu and Yamamoto, 2015). The presence of ample fluid (gas, vapor, magma, or their mixtures) in volcanic edifices introduces an additional class of vibration/wave phenomena to the well-studied elastic formulations. There exists a class of waves in the two-phase system of a solid–liquid composite that travels slower than any of the sound velocities of the pure material constituting the two-phase system (i.e., the wave speed can be slower than that of the liquid phase). Rayleigh wave and Stoneley wave, as well as the crack wave found by Chouet (1986), may be included in this class. The simplest example of slow waves may be the *P*-wave propagation in a medium that consists of a fluid layer sandwiched by two elastic half-spaces (Ferrazzini and Aki, 1987).

Fig. 15.9, taken from Kumagai (2009), illustrates the relationship between the phase velocities of the acoustic and crack waves (Fig. 15.9B), generated in a solid–fluid composite. In this figure, a compressive fluid layer is sandwiched between two elastic half-spaces, which mimics a dyke containing basaltic magma with a gas-volume fraction of 15% (Fig. 15.9A). The dispersion relationship in Fig. 15.9B clearly indicates that the phase velocity of the fluid–solid coupled crack wave depends on the wavelength. Kumagai (2009) showed that the normal displacement, symmetric with respect to  $x_3$ , causes the deformation of the fluid–solid boundary. The deformation has the effect of lowering the phase velocity of the crack wave.

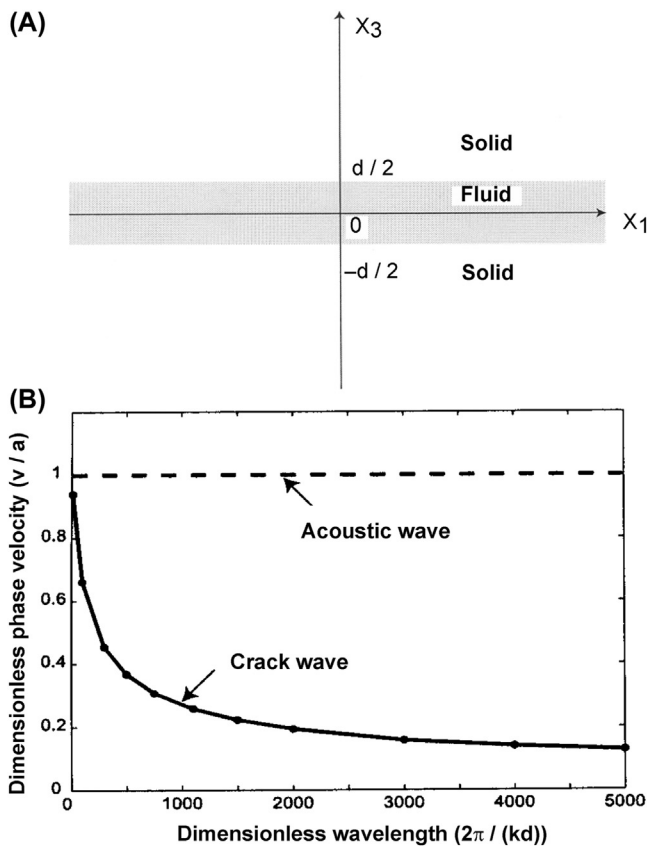


**FIGURE 15.9** (A) Fluid layer of thickness  $d$  sandwiched between two elastic half-spaces bounded at  $x_3 = -d/2$  and  $x_3 = d/2$  and (B) the phase velocities of the acoustic wave (dashed line) and crack wave (solid line) generating within this fluid–solid composite. These are normalized by the fluid acoustic velocity, and plotted as a function of the wavelength normalized by the thickness of the fluid layer. From Kumagai, H., 2009. *Volcano seismic signals, source quantification of*. In: R.A. Meyers (Ed.), *Encyclopedia of Complexity and Systems Science*. Springer-Verlag, Berlin, pp. 9899–9932.

### 15.1.4.2 Source of Long-Period and Very-Long-Period Seismic Signals

Waveform inversions of LP and VLP seismic signals show that their sources may be modeled by the superposition of six moment-tensor and three single-force components (Kumagai et al., 2005; Ohminato et al., 1998; Waite et al., 2008; Chouet, 2009). The modeling is applied at a common point within the volcano, and therefore, as noted by Aster et al. (2008), is a highly idealized representation of a spatially distributed source. However, this point approximation may be good enough when scaled to the wavelength of LP seismic waves.

An example of the source-time functions obtained for the July 2, 2005, LP event recorded during lava dome growth at Mount St. Helens (Waite et al., 2008) is shown in Fig. 15.10. The moment tensor is dominated by the

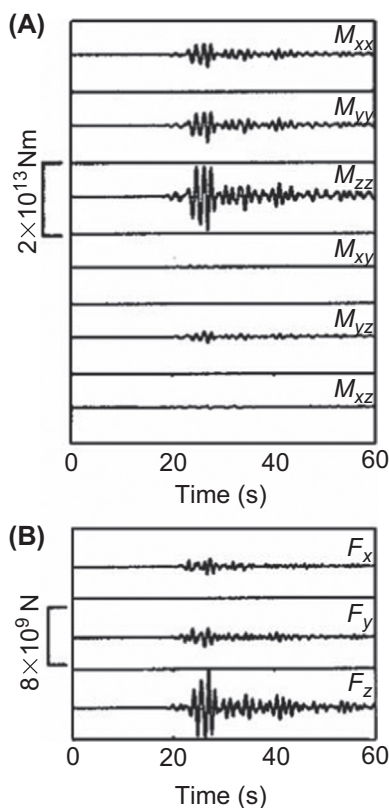


**FIGURE 15.10** Source-time function of the July 2, 2005 long-period (LP) event recorded at Mount St. Helens. It is dominated by the volumetric components of the moment tensor (A) and a vertical single force (B). From Waite, G.P., Chouet, B.A., Dawson, P.B., 2008. Eruption dynamics at Mount St. Helens imaged from broadband seismic waveforms: interaction of the shallow magmatic and hydrothermal systems. *Journal of Geophysical Research* 113, B02305.

three diagonal terms (dipole components) of the seismic moment tensors ( $M_{xx}$ ,  $M_{yy}$ , and  $M_{zz}$ ) and a vertical single-force component  $F_z$ . The dipole components are in phase and there is very little energy on the shear components, indicating that the mechanism represents a volumetric source. The amplitude ratio  $F_z/M_{zz}$  is roughly  $2.8 \times 10^{-4}/\text{m}$ . The moment-tensor and single-force components contribute equally to the waveform amplitudes when this ratio is  $10^{-3}/\text{m}$  (Ohminato et al., 1998). Therefore the  $F_z$  contribution is equal to 28% of the wave amplitudes. Waite et al. (2008) consider that the vertical-force component may be explained by oscillations of the growing spine and possibly the entire dome.

An example of the source-time functions obtained for the Popocatepetl VLP source during the May 23, 2000 Vulcanian degassing (Chouet et al., 2005) is shown in Fig. 15.11. Volumetric components of the moment tensor clearly dominate the solution, with the source-time functions characterized by almost identical time histories. Accompanying the volumetric source components are two clear components of a single force in the east and vertical directions contributing up to 16% and 13% of the wave amplitude, respectively.

This dominance of the components of the moment tensor above the single-force components is a characteristic of the majority of LP and VLP sources obtained from an inversion of the seismic signals.

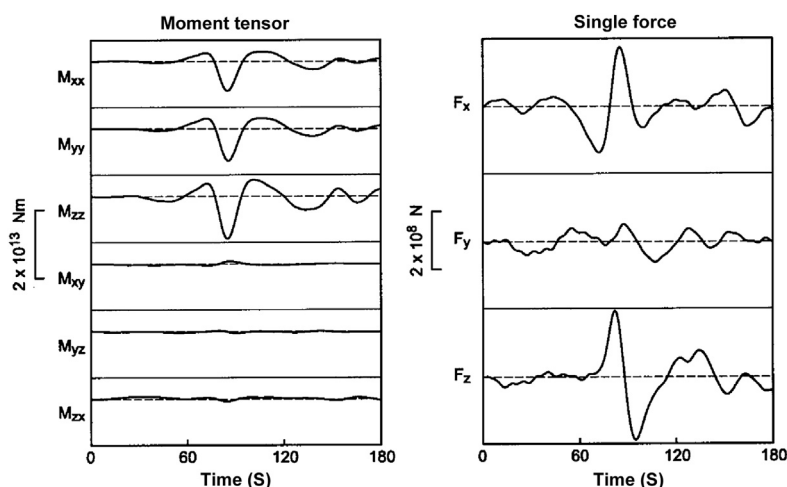


**FIGURE 15.11** Source-time function of the May 23, 2000 very-long-period (VLP) event recorded at Popocatepetl volcano. It is dominated by the volumetric components of the moment tensor and the eastern and vertical components of a single force. From Chouet, B., Dawson, P., Arciniega-Ceballos, A., 2005. Source mechanism of Vulcanian degassing at Popocatepetl volcano, México, determined from waveform inversions of very long period signals. *Journal of Geophysical Research* 110, B07301.

**TABLE 15.1** Ratios of Three Diagonal Terms of the Seismic Moment Tensors  $M_{11}$ ,  $M_{22}$ , and  $M_{33}$  for Some Non–Double-Couple Sources of Volcanic Earthquakes (the Elastic Modulus are Taken as  $\mu = \lambda$ ).

| Type of source           | $M_{11}:M_{22}:M_{33}$ | References                                    |
|--------------------------|------------------------|-----------------------------------------------|
| Spherical                | 1:1:1                  | <a href="#">Kumagai (2009)</a>                |
| Vertical cylinder        | 2:2:1                  | <a href="#">Kumagai (2009)</a>                |
| Vertical tensile crack   | 3:1:1                  | <a href="#">Kawakatsu and Yamamoto (2015)</a> |
| Horizontal tensile crack | 1:1:3                  | <a href="#">Kumagai (2009)</a>                |

## 15.2 GEOMETRY OF THE SOURCES OF LONG-PERIOD AND VERY-LONG-PERIOD SEISMIC SIGNALS

LP and VLP seismic signals originating from the vibration of the stable and nondestructive fluid-filled sources (dykes, conduits, cracks) carry information about the geometry of these sources. Inversion of the seismic waves of the LP and VLP signals gives the seismic moment tensor of the source. As was mentioned in Section 3.1.3, the ratios of three diagonal terms of the seismic moment tensors  $M_{11}$ ,  $M_{22}$ , and  $M_{33}$  (or  $M_{xx}$ ,  $M_{yy}$ , and  $M_{zz}$ ) allow the discrimination of the geometry of the non–double-couple sources of seismic waves indicating a crack, a cylindrical pipe, or a sphere ([Table 15.1](#)).

The source of the February 1, 1996 LP event, which occurred just 12 hours before the volcanic crisis at Kilauea volcano, was modeled as a horizontal crack ([Kumagai et al., 2005](#)). The source-time functions of the six moment-tensor components calculated from the waveform inversion are shown in [Fig. 15.12](#). Domination of the  $M_{zz}$  component over the  $M_{xx}$  and  $M_{yy}$  components is seen ( $M_{11}:M_{22}:M_{33} \cong 1:1:3$ ) reflecting a horizontal crack source.

A cylindrical pipe was obtained as a model of the source of LP signals for the October 30, 1986 event associated with the explosion at Sakurajima volcano ([Uhira and Takeo, 1994](#)). The source-time functions of the six moment-tensor components calculated from the waveform inversion are shown in [Fig. 15.13](#). The amplitude ratios  $M_{xx}:M_{yy}:M_{zz}$  are about 1.8:1.8:1. This nearly corresponds to a contraction of a cylindrical region almost vertically embedded in an elastic medium, equivalent to the conduit.

Sometimes the moment-tensor solutions give the ratios of three diagonal terms of the seismic moment tensors  $M_{11}$ ,  $M_{22}$ , and  $M_{33}$ , which differ from the simple situations shown earlier. For example, the ratios of three diagonal terms for the 2000 Vulcanian degassing event at Popocatepetl were 1.20:1.45:2.00 (see [Fig. 15.11](#)), which, as considered by [Chouet et al. \(2005\)](#), indicates the possibility of a dual mechanism consisting of two elements: a shallow dipping crack (sill) and an NE trending quasivertical crack (dyke).

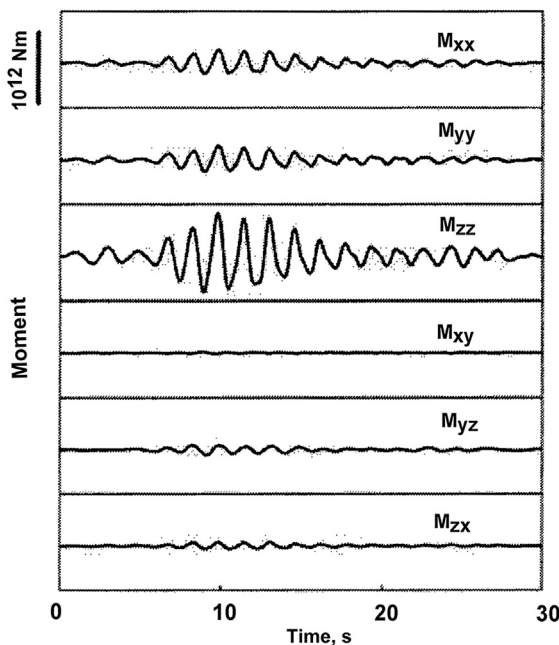


**FIGURE 15.12** Source-time functions of the six moment-tensor components obtained from wave-form inversion of the long-period (LP) waveforms recorded at eight stations on February 1, 1996, at Kilauea volcano. From Kumagai, H., Chouet, B.A., Dawson, P.B., 2005. Source process of a long-period event at Kilauea volcano, Hawaii. *Geophysical Journal International* 161, 243–254.

### 15.3 TYPE OF FLUID WITHIN THE FLUID-FILLED CRACKS

The harmonic oscillatory signatures of the LP events are interpreted as oscillations of fluid-filled resonators at the source of these signals. The type of fluid filling these resonator systems may be identified from the comparison of the complex frequencies of decaying harmonic oscillations in the tail of the seismogram of the LP seismic signals and the model crack data.

#### 15.3.1 Crack Model

The rectangular crack model with a width  $W$ , length  $L$ , and aperture of the crack  $d$  was used for the simulation of the dynamics of a fluid-filled crack (see Section 11.3.1, Image 11.9 from Chapter 11). This model, proposed by Chouet (1986), is characterized by the independent parameters  $\alpha/a_p$  and  $\rho_f/\rho_s$ , where  $a$  is the speed of sound propagating within the fluid in the crack and  $\alpha_p$  is the  $P$ -wave velocity of the rock matrix, and  $\rho_f$  and  $\rho_s$  are the densities of the fluid and rock matrix, respectively. The waveform, obtained from the simulation of the crack excitation in the form of a pressure step applied over a small area of the crack wall at the center of the crack, can be regarded as the velocity impulse response of the crack.

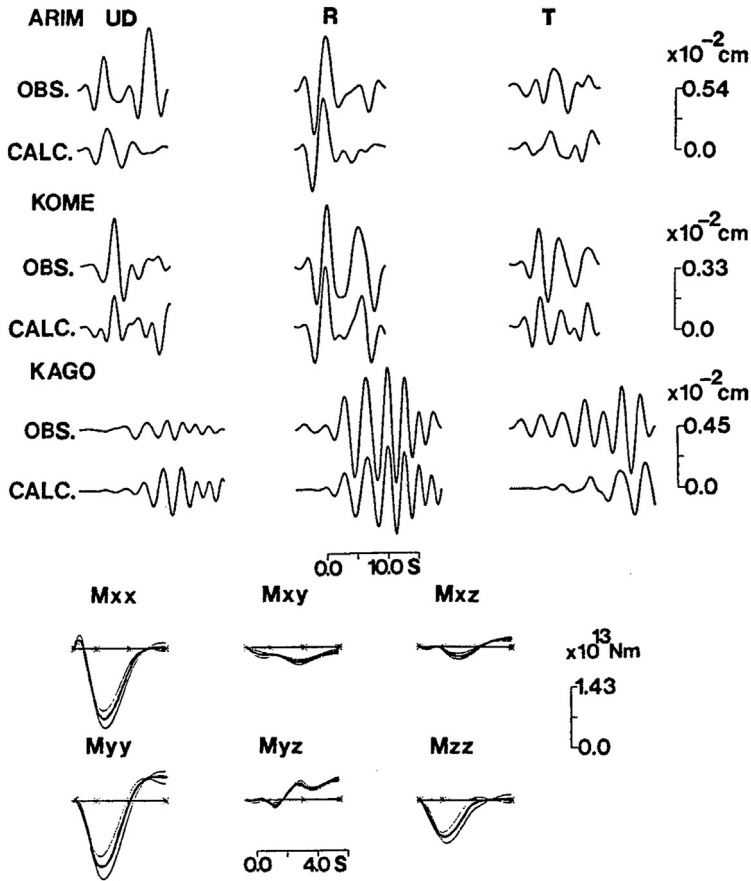


**FIGURE 15.13** The result of inversion of the long-period (LP) waveforms recorded on October 30, 1986, at Sakurajima volcano at three stations (ARIM, KOME, and KAGO) and the estimated source-time functions of the six moment-tensor components. Thin lines show the error bound of the estimated time histories. The observed and calculated seismograms are abbreviated as OBS and CALC. The vertical, radial, and transverse components are denoted by UD, R, and T, respectively. From Uhira, H., Takeo, M., 1994. The source of explosive eruptions of Sakurajima volcano, Japan. *Journal of Geophysical Research* 99, 17775–17789.

### 15.3.2 Complex Frequencies of the Long-Period Seismic Signal for Different Fluids

Complex frequencies refer to the frequencies and quality  $Q$  values of decaying harmonic oscillations of an LP signal. The complex frequency is defined as  $f - ig$ , where  $f$  is the frequency,  $g$  is the growth rate, and  $i = \sqrt{-1}$ . Accordingly, the factor  $Q$  of the signal is defined as  $-f/2g$ . The crack model simulations predict  $Q$  due to the radiation loss, or  $Q = Q_r$  (Kumagai and Chouet, 2000).



Fig. 15.14 demonstrates the contour diagrams of  $Q_r$  and the dimensionless frequency  $v = fL/\alpha$  and plots of  $\alpha/a$  versus  $\rho_f/\rho_s$  curves for various types of fluids calculated for the synthetic waveforms obtained from the simulation of the crack excitation (Fig. 15.15). As a result, it is possible to discriminate

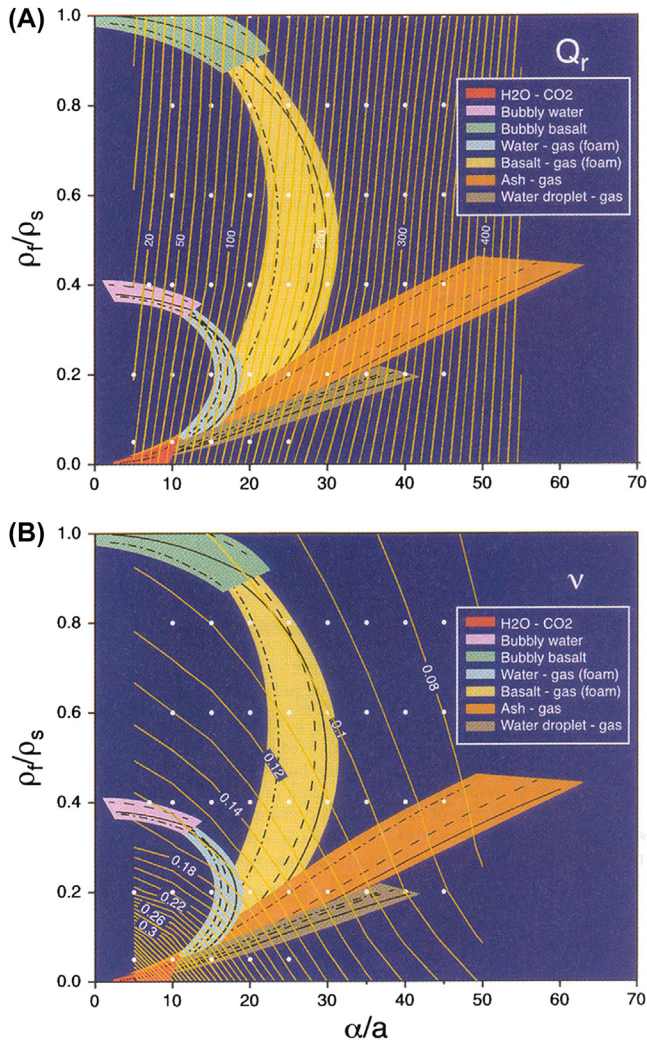


**FIGURE 15.14** Contour diagrams of  $Q_r$  (A) and dimensionless frequency  $v = fL/\alpha$  and plots of  $\alpha/a$  (B) versus  $\rho_f/\rho_s$  curves for various types of fluids. Dots indicate points where the synthetic waveforms are calculated. Dashed, solid, and dash-dotted lines indicate  $\alpha/a$  and  $\rho_f/\rho_s$  calculated for fluids at pressures of 5, 10, and 20 or 25 MPa, respectively. From Kumagai, H., Chouet, B., 2000. Acoustic properties of a crack containing magmatic or hydrothermal fluids. *Journal of Geophysical Research* 105, 25493–25512.

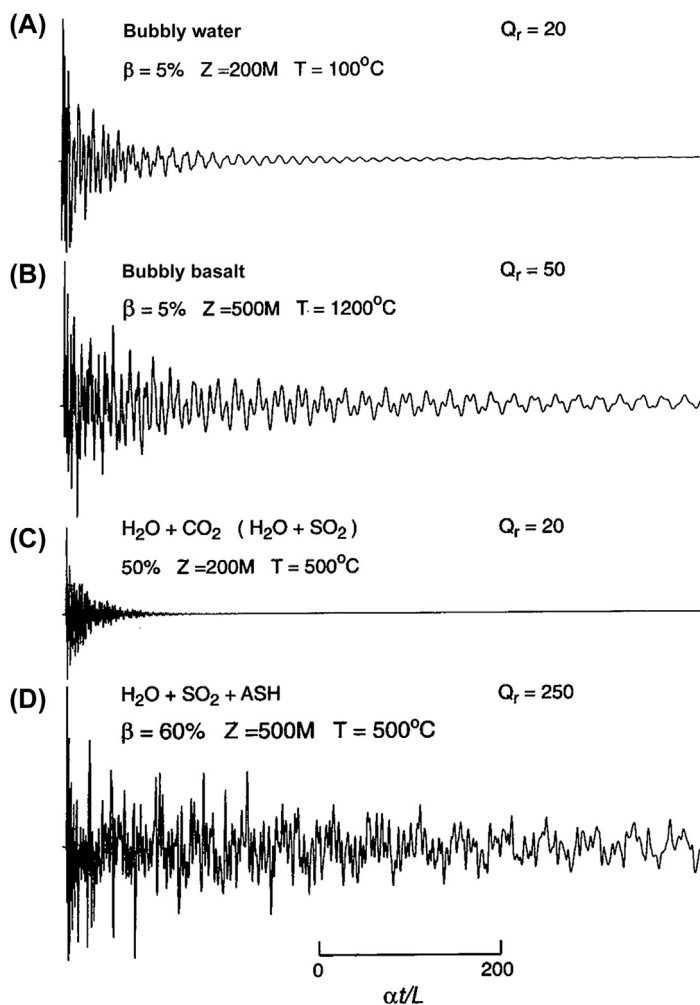


**FIGURE 15.15** Synthetic far-field compressional waveforms produced by the excitation of a crack containing different types of fluids. The crack is embedded in a rock with fixed properties  $\alpha = 2700$  m/s,  $\rho_s = 2500$  kg/m<sup>3</sup>. (A) Waveform for a crack at a depth of 200 m containing bubbly water with a 5% gas-volume fraction at a temperature of 100°C. (B) Waveform for a crack at a depth of 500 m containing a bubbly basalt with a 5% gas-volume fraction at a temperature of 1200°C. (C) Waveform for a crack at a depth of 200 m containing a 50% mixture of either H<sub>2</sub>O–CO<sub>2</sub> or H<sub>2</sub>O–SO<sub>2</sub> gases at a temperature of 500°C. (D) Waveform for a crack at a depth of 500 m containing an ash-laden gas with a gas-volume fraction of 60% at a temperature of 500°C. From Chouet, B., 2003. *Volcano seismology. Pure and Applied Geophysics* 160, 739–788.

various mixtures of gas, liquid, and solid, including gas–gas mixtures (H<sub>2</sub>O–CO<sub>2</sub> and H<sub>2</sub>O–SO<sub>2</sub>), liquid–gas mixtures (water–H<sub>2</sub>O and basalt–H<sub>2</sub>O), and dusty and misty gases (ash–SO<sub>2</sub> and water droplet–H<sub>2</sub>O; Kumagai and Chouet, 2000).

### 15.3.3 Identification of the Type of Fluid from Long-Period Seismic Signals

To estimate the complex frequencies  $f$ , a Sompi analysis (Hori et al., 1989), based on an autoregressive model and assuming harmonic decaying oscillations, may be applied. In this method of spectral analysis, a signal is deconvolved into a linear combination of coherent oscillations with decaying amplitudes (called wave elements) and additional noise. An amplitude value is also calculated for each wave element. Generally, to represent a set of complex frequencies, they are plotted on a 2D plane with the frequency  $f$  and growth rate  $g$  axes, called the “ $f$ – $g$  diagram.” Wave elements scattering widely in the plot are considered to be noise, while wave elements densely populated on the theoretical frequency lines are considered to be the dominant spectral components of the signal (Hori et al., 1989).

Fig. 15.16A shows the LP seismic signal recorded near Kilauea volcano, Hawaii (Kumagai et al., 2005). To determine the complex frequencies from the waveform, the onset portion of the LP seismic signal, which contains the inhomogeneous part of the signal, was deleted. The Fourier spectrum of this signal (Fig. 15.16B) was characterized by two dominant peaks near 0.6 and 1.3 Hz. The  $Q$  values of the two peaks are in the range 10–20 (Fig. 15.16C). Comparison with the diagrams of Fig. 15.14 gives a bubbly water fluid filling the crack.

The Fourier and Sompi frequency domain analyses were applied to the tornillo-type LP events (Fig. 15.17A) recorded at Vulcano volcano, Italy during 2007–08 (Milluzzo et al., 2010). The volcanic activity of this volcano, since the last eruption in 1888–90, produces fumarolic emissions. The Fourier analysis showed that most of the tornillo signals are characterized by two sharp spectral peaks at frequencies of about 6 and 10 Hz (Fig. 15.17C). To obtain information about the decay of the tornillo signals, the Sompi analysis was applied. The amplitude information was added by using a color scale for the dots (Fig. 15.17D). Such amplitude information is useful for further discrimination of the noise and dominant spectral components of the signal. The former and the latter are generally characterized by small and large amplitudes, respectively (Hori et al., 1989). A 10-second window from the recording of the vertical component of short-period seismic station IVCR, situated at a distance of about 2 km from the crater, was analyzed for each tornillo event. The signal was filtered in two frequency bands, 5–8 and 8–12 Hz, and the Sompi method was applied to each band. The frequency values obtained confirm the results of the Fast Fourier Transformation analysis. It was noted that most of the quality factor values ranged between 20 and 400. According to Fig. 15.14, the fluid-filled resonator could be an ash–gas or water droplets–gas mixture.

A swarm of LP seismic events was recorded in December 2001 before heightened eruptive activity at Tungurahua volcano, Ecuador (Molina et al., 2004). The waveforms of two of them are shown in Fig. 15.18A.

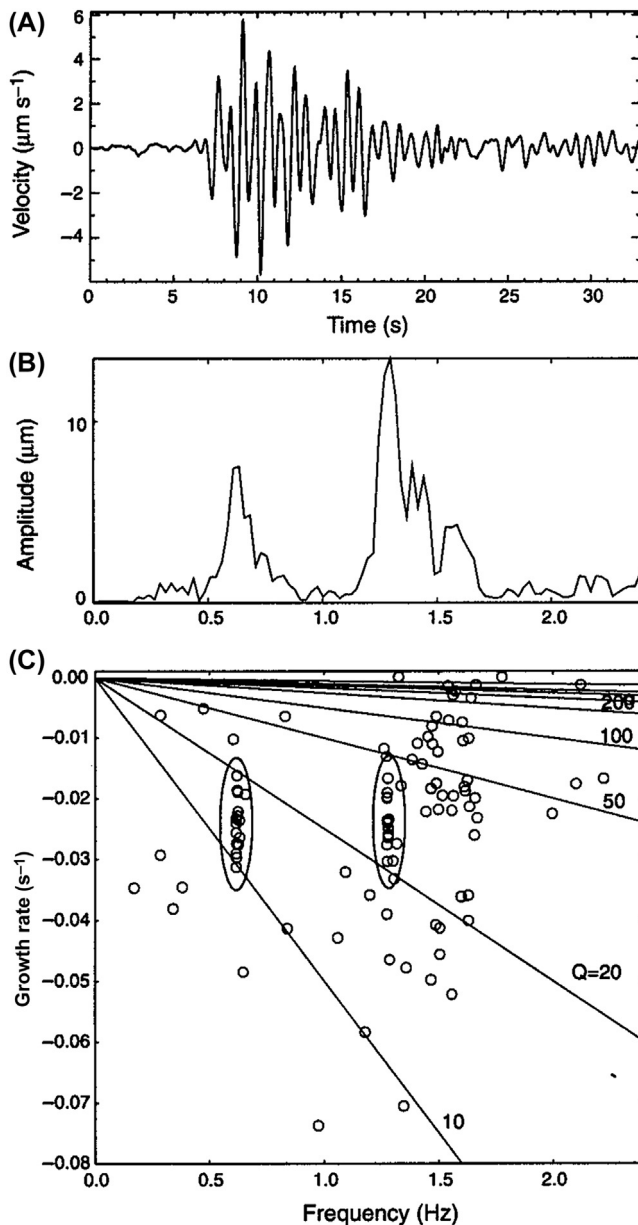


**FIGURE 15.16** Seismic record of the February 1, 1966 long-period (LP) event recorded at Kilauea volcano (A), its Fourier amplitude spectrum (B) and the plots of the complex frequencies estimated for this waveform (C). The clusters of points within ellipses represent clear signals. The solid lines represent lines along which the  $Q$  factor is constant. *From Kumagai, H., Chouet, B.A., Dawson, P.B., 2005. Source process of a long-period event at Kilauea volcano, Hawaii. Geophysical Journal International 161, 243–254.*

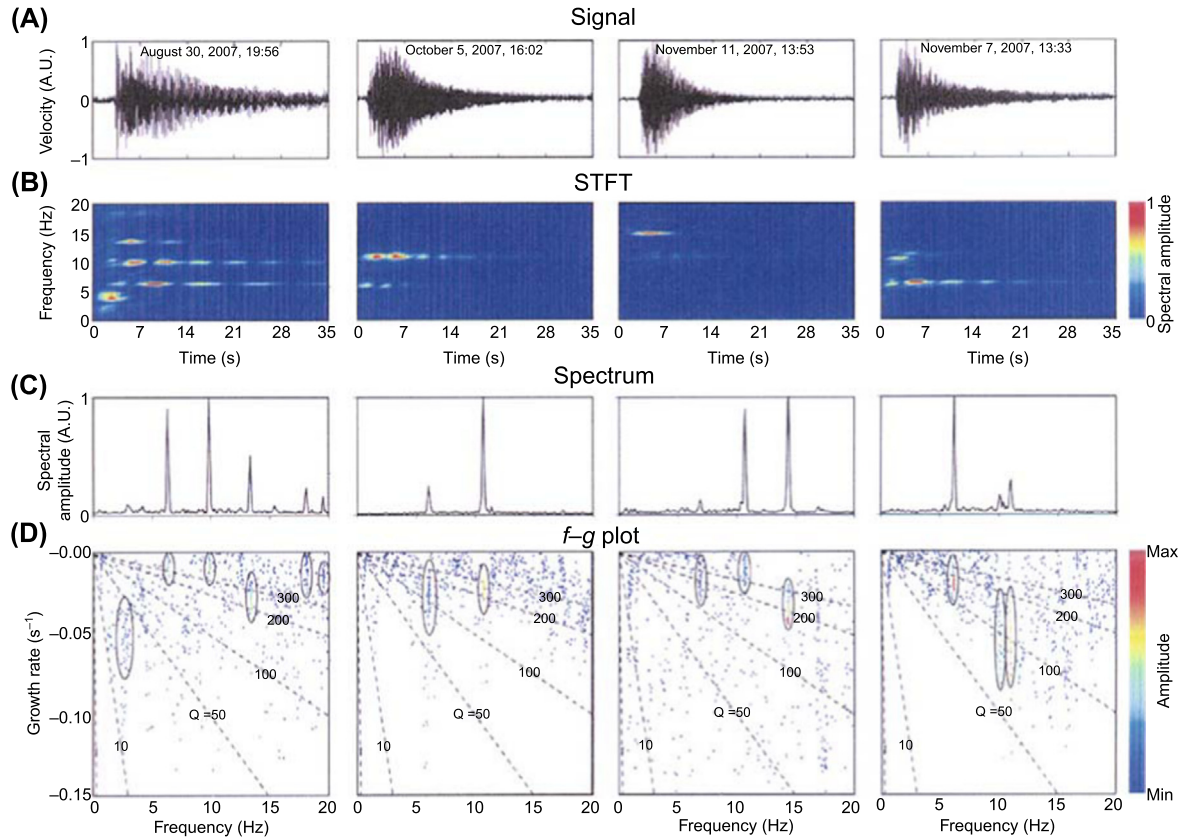


**FIGURE 15.17** Vertical component waveforms (A), short-time Fourier transform spectrogram (STFT) (B), Fourier spectra (C), and  $f$ - $g$  plots (D) of four examples of tornillos recorded at Vulcano volcano during 2007–08. The color of the dots in (D) indicates the amplitude values of the wave elements (see color scale). Clusters of dots in (D) (*encircled with solid ellipses*) indicate dominant spectral components of the signal, and scattered dots represent noise. *From Milluzzo, V., Cannata, A., Alparone, S., Gambino, S., Hellweg, M., Montalto P., et al., 2010. Tornillos at Vulcano: clues to the dynamics of the hydrothermal system. Journal of Volcanology and Geothermal Research 198, 377–393.*

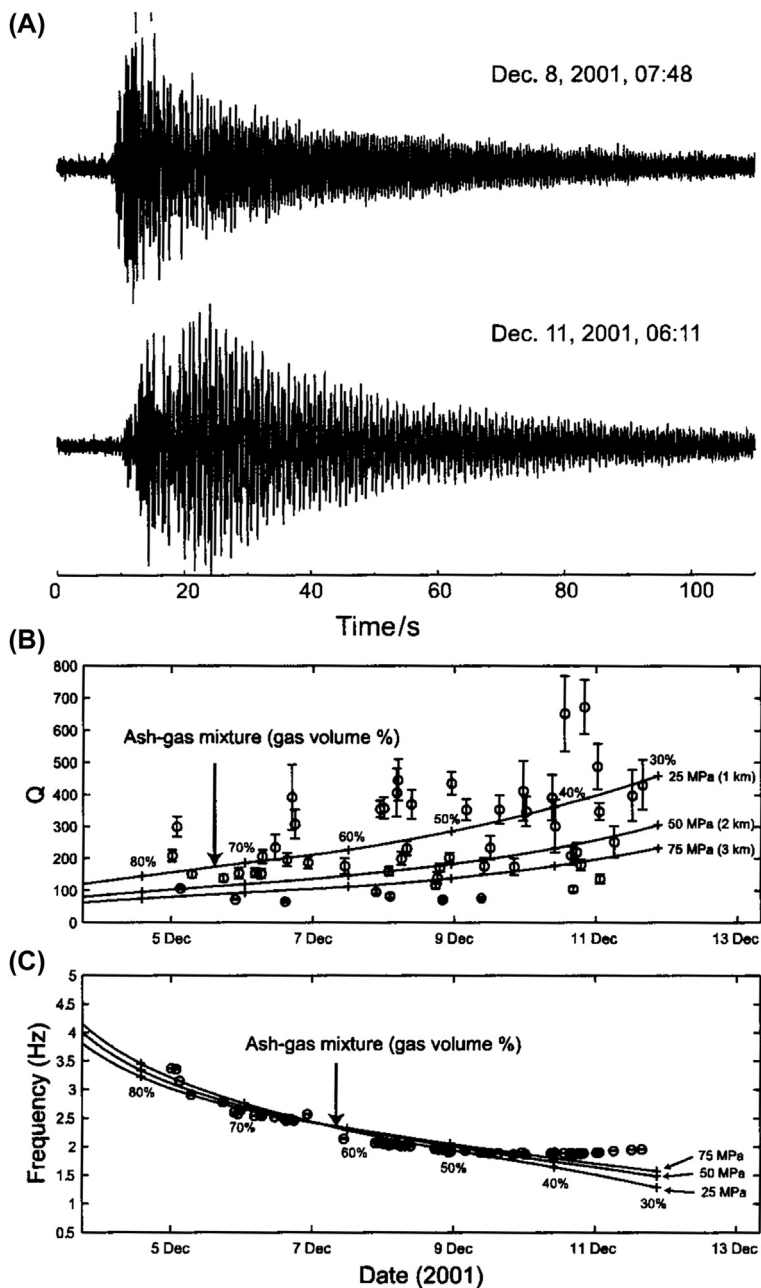


**FIGURE 15.18** Seismic records of the December 2, 2001 long-period (LP) events recorded at Tungurahua volcano (A) and the temporal variations in  $Q$  (B) and frequency (C) determined for LP events recorded at Tungurahua volcano during the December 2001 seismic swarm. The *solid lines* show the best fits of frequency and  $Q$  calculated for a crack containing the ash–gas mixture at various pressures between 25 and 75 MPa corresponding to the source depths between 1 and 3 km below the summit, in which crack lengths of 200, 230, and 240 m at depths of 1, 2, and 3 km, respectively, were assumed. *After Molina, I., Kumagai, H., Yepes, H., 2004. Resonances of a volcanic conduit triggered by repetitive injections of an ash-laden gas. Geophysical Research Letters 31, L03603.*

Sompi analysis was applied to estimate the complex frequencies and  $Q$  for 57 LP events. Fig. 15.18B and C shows that both the frequency and  $Q$  systematically changed during the swarm. The factor  $Q$  gradually increased from 100 to 400, while the frequency gradually decreased from 3.5 to 3 Hz. Molina et al. (2004) considered that the observed temporal variations may be explained by a gradual increase in the ash content in the crack at the source of the LP events.

## 15.4 LOCATION OF THE SOURCES OF LONG-PERIOD AND VERY-LONG-PERIOD EVENTS

This type of seismicity cannot be located by conventional means due to the lack of impulsive body wave phases in the records. The source locations of LP and VLP events are obtained in a probabilistic sense using methods such as frequency-slowness analysis (Almendros et al., 2002), waveform inversion (Chouet et al., 2003), cross-correlation (De Barros et al., 2009), and waveform semblance (Almendros et al., 2002). The majority of the sources, generating LP and VLP signals at volcanoes, have been located within the depth interval between 100 and 400 m beneath the active crater or dome (Merapi, Hidayat et al., 2002; Stromboli, Chouet et al., 2003; Mount St. Helens, Waite et al., 2008; Kilauea, Kumagai et al., 2005; Etna, De Barros et al., 2009); however, some of them were located much deeper, between 1 and 8 km (Cotopaxi, Molina et al., 2008; Popocatepetl, Chouet et al., 2005; Aso, Legrand et al., 2000; Kilauea, Almendros et al., 2002).

Battaglia et al. (2003) developed a methodology for locating LP events using seismic amplitudes instead of using arrival times. The method is based on seismic amplitudes corrected for station site effects using coda site amplification factors. Once corrected, the spatial distribution of amplitudes shows smooth and simple contours for the LP events. The density of the contours also gives an estimation of the depth of the source. On the basis of the simplicity of these distributions, the inversion methods for locating their origins were developed. To achieve this, the decrease of the amplitude as a function of the distance to the source was approximated by the decay of either surface or body waves in a homogeneous medium. Amplitudes at each station were calculated using root mean square on 20.28-second windows starting shortly before the onset of the signal.

Fig. 15.19A shows the location of the LP event occurring at Kilauea volcano in January 1998 using seismic amplitudes (Battaglia et al., 2003). The best source is plotted as a white star on the plane projection showing the summit of the volcano, as well as on the EW and NS cross-sections. This methodology was applied to the location of totally 2896 LP events occurring from January 1997 to December 1999 below Kilauea (Fig. 15.19B). The foci of LP events clustered at two depth intervals, above 1 km below sea level (b.s.l.; shallow events) and between 3 and 10 km b.s.l. (deep events).



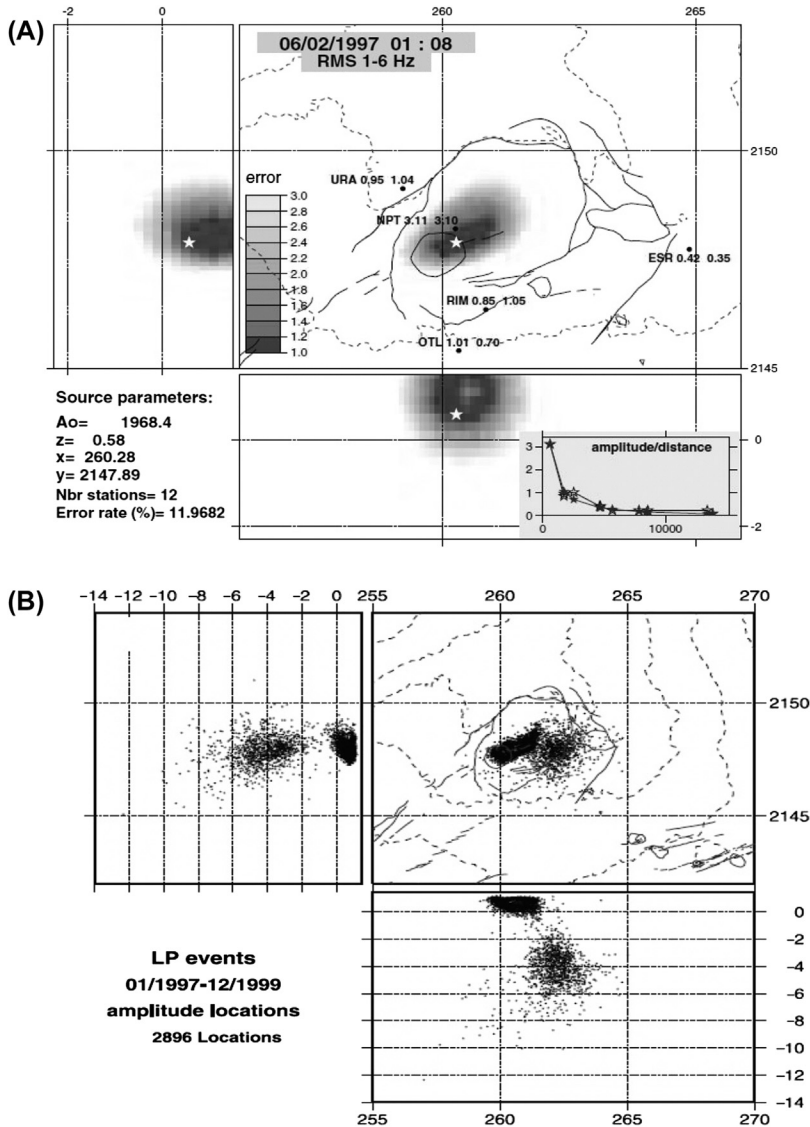


**FIGURE 15.19** Location of long-period (LP) events, occurring during January 1998 below Kilauea volcano, using seismic amplitudes. (A) Location of an LP event and the EW and NS cross-sections. (B) Location of 94 LP events and the EW and NS cross-sections. The decays (observed and modeled) of the amplitude as a function of the distance are shown in the inset. In (B), the routine catalog locations are plotted in the background as gray points. See the text for more explanations. After Battaglia, J., Got, J.-L., Okubo P.G., 2003. Location of long-period events below Kilauea using seismic amplitudes and accurate relative relocation. *Journal of Geophysical Research* 108, 2553.



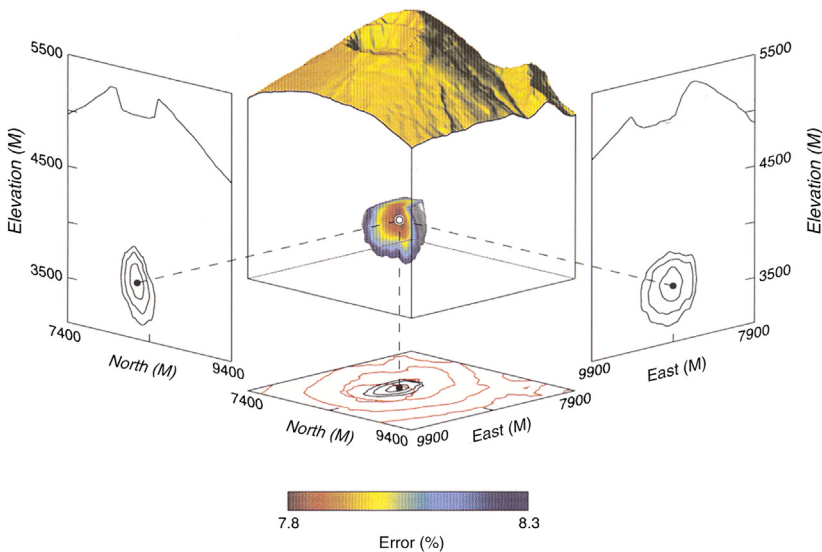


**FIGURE 15.20** Source location (white dot) of the very-long-period (VLP) event occurring during the May 23, 2000 explosive activity at Popocatepetl volcano from waveform inversion. A southwest-looking cutaway view of the volcano is shown at the top to help identify the location of the VLP source within the volcanic edifice. Contours represent 7.9%, 8.1%, and 8.3% errors. From Chouet, B., Dawson, P., Arciniega-Ceballos, A., 2005. Source mechanism of Vulcanian degassing at Popocatepetl volcano, México, determined from waveform inversions of very long period signals. *Journal of Geophysical Research* 110, B07301.

Fig. 15.20 shows the location of the source centroid obtained by waveform inversion of VLP seismic signals generated during the May 23, 2000 explosive activity at Popocatepetl volcano, Mexico. A network consisting of 15 broadband seismic stations was used. The source centroid that best fit the data was located at an elevation of 3600 m, approximately 1500 m below the western perimeter of the crater floor (Chouet et al., 2005).

The methodology of cross-correlating signals, obtained from 50 broadband seismic stations, was used to locate the LP seismic events at Etna volcano during June 2008 (De Barros et al., 2009). To extract the LP events, the short-term average/long-term average method (2 seconds over 20 seconds window lengths, with a threshold of 2.5) on the bandpass filtered data (0.2–1.5 Hz) was used. Selected events were classified using a cross-correlation analysis between all pairs of signals. The events that gave a correlation coefficient greater than 0.9 with all events on at least three out of the four permanent stations close to the summit formed two families with a similar number of events (63 and 66, respectively). The location of the LP events was computed in two steps: First, the mean position for each family was found, and second, the individual events were located using this first position.

Fig. 15.21 shows the locations of 129 sources with dominant frequencies, ranging between 0.3 and 1.2 Hz, which generated two families of LP seismic

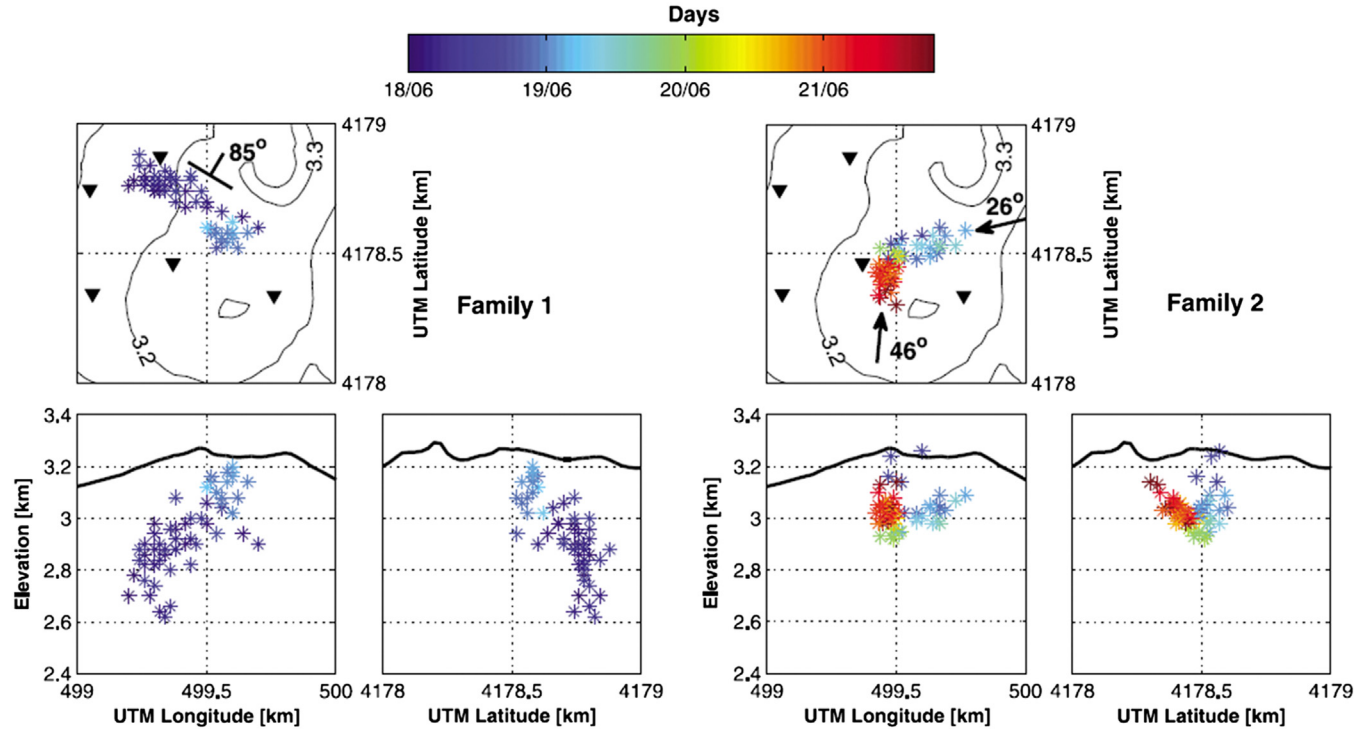


**FIGURE 15.21** Location for the sources of family 1 (left) and family 2 (right), generating long-period (LP) seismic signals at Etna, performed by cross-correlating signals for pairs of stations, with colors indicating temporal evolution. Color scale (days) is common for both families. Views are from above, south and west. Triangles are some of the broadband stations. “⊥” and “→” symbols represent the planar and pipe-shape structures, with dip and plunge angles, respectively. *From De Barros, L., Bean, C.J., Lokmer, I., Gilberto Saccorotti, G., Zuccarello, L., Gareth S. O’Brien, G.S., et al., 2009. Source geometry from exceptionally high resolution long period event observations at Mt Etna during the 2008 eruption. Geophysical Research Letters 36, L24305.*

signals. The events for both families are shallow: from 50 to 800 m for family 1 and from 20 to 400 m for family 2. The epicenters of the events are close to the summit craters and are slightly different for both families.

The location of the sources of VLP seismic signals at Kilauea volcano during the period September 9, 1999, to September 16, 1999, was performed using a method based on analysis of semblance (Almendros et al., 2002). Semblance is a measure of the similarity of multichannel data weighted according to the rectilinearity of the particle motions. This method determines the point for which the radial components of the signals recorded at different receivers, aligned to correct for the wave propagation delay, show maximum similarity. The procedure, used to obtain a source location by the semblance method, consists of three steps: (1) determination of the spatial extent of the region of study by defining a three-dimensional grid of assumed source positions; (2) fixation of the interval of signal considered by selecting the start time of the analysis window and its duration; and (3) calculation of semblance using the arrival times calculated for each node of the source grid. This procedure gives a three-dimensional distribution of radial semblance.

Two source regions for 16 VLP events were identified (Fig. 15.22). One of them was located at a depth of about 1 km beneath the northeastern edge of the Halemaumau pit crater. The second region was located at a depth of about 8 km below the northwest quadrant of Kilauea caldera (Almendros et al., 2002).

## 15.5 CONCEPTUAL MODELS OF THE RELATIONSHIP BETWEEN THE SOURCES OF THE LONG-PERIOD AND VERY-LONG-PERIOD SEISMIC SIGNALS AND THEIR ROLE IN ERUPTIVE PROCESS

A study of the sources of LP and VLP seismic signals recorded during the period from November 2003 to May 2006 at Etna using radial semblance showed that these two types of signal are not always generated by a common source but by different sources connected to each other. It was obtained that a deeper dyke, at about 2.5 km above sea level (a.s.l.), was the source of the VLP events, while a shallower crack, situated 500 m above the dyke, generated LP signals (Cannata et al., 2009). The same relationship between depth and the generation of LP and VLP seismic signals was reported for the Cotopaxi and Mount St. Helens volcanoes (Molina et al., 2008; Waite et al., 2008).

The relative source process for the July 2, 2005 Mount St. Helens LP and VLP events, located using waveform inversion, was discussed by Waite et al. (2008). In the seismic record, the main part of the VLP signal occurred after the initiation of the LP event. The positions of the LP and VLP sources are shown in Fig. 15.23. The LP source is very shallow, at about 2000 m elevation and directly beneath the growing dome. The VLP source is 250 m deeper and 400 m NNW, beneath the SW end of the 1980s lava dome. The LP source does not seem to be within the eruption conduit.

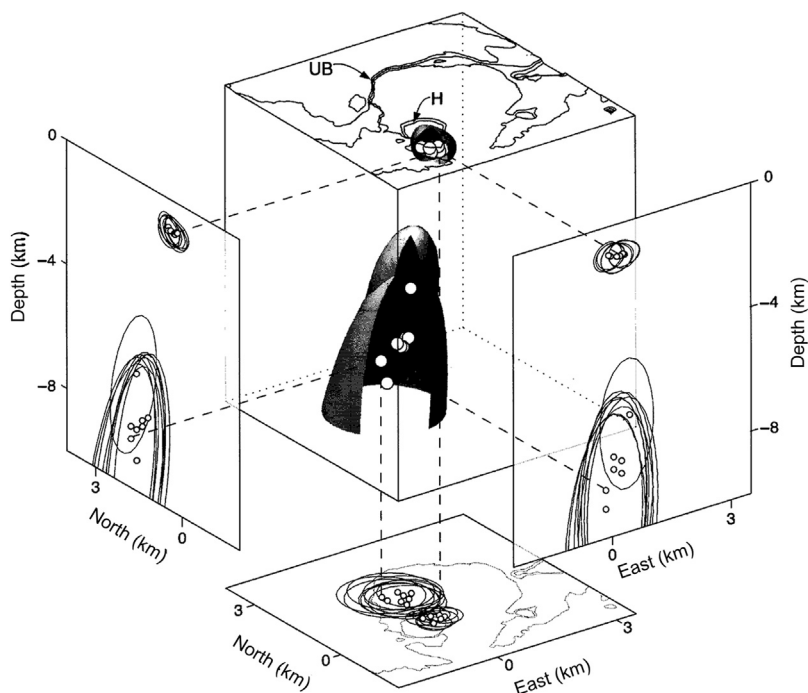


**FIGURE 15.22** Source locations of the 16 very-long-period (VLP) events recorded at Kilauea volcano during September 1999 and located by a method based on analysis of semblance and particle motion. The sources are shown by large open circles. The topography of the Kilauea caldera is shown at the top of the figure. Contours on the side planes are projections of the error regions for the source locations. The gray surfaces represent the outline of the error regions for all events. H, Halemaumau; UB, uwēkahuna bluff. From Almendros, J., Chouet, B., Dawson, P., Bond, T., 2002. Identifying elements of the plumbing system beneath Kilauea volcano, Hawaii, from the source locations of very-long-period signals. *Geophysical Journal International* 148, 303–312.

A possible scenario for the LP events at Mount St. Helens involves the repeated pressurization of a subhorizontal crack that lies beneath the southern crater and new lava dome. Steam generated by the heat of groundwater, in addition to exsolved steam from the magma, fed into the crack, gradually expanding and pressurizing it. The southern, higher altitude end of the crack must be sealed to prevent steam from escaping. When the pressure reaches its threshold, some steam is forced out of the crack through fractures that eventually lead to the surface or glacier–crater floor interface. Waveform inversion of the VLP events revealed a north-dipping sill plus a near-vertical dyke or pipe. The VLP source sill was roughly in the plane of the LP source crack. The main VLP signal was generated by deflation–inflation of the sill and near-simultaneous inflation–deflation of the dyke. Large, subhorizontal forces that accompanied the crack volume changes were interpreted as reaction forces

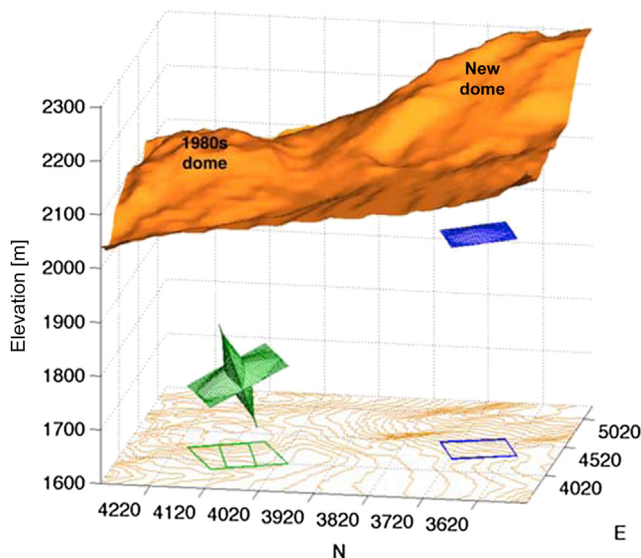


**FIGURE 15.23** Relative position of the sources that generated the July 2 2005 long-period (LP) and very-long-period (VLP) seismic signals during lava dome growth in the crater of Mount St. Helens. The east–northeast, three-dimensional, perspective view of the LP crack (blue [dark gray in print versions]) and VLP sill-dyke (green [light gray in print versions]) sources are shown. Topography is shown at the bottom as a surface and is contoured at 20 m. Distances are in meters. From Waite, G.P., Chouet, B.A., Dawson, P.B., 2008. Eruption dynamics at Mount St. Helens imaged from broadband seismic waveforms: interaction of the shallow magmatic and hydrothermal systems. *Journal of Geophysical Research* 113, B02305.

on the Earth resulting from transient mass advection in the conduit. It was proposed that the collapse of the LP crack and related sagging of the dome triggers a response in the magma conduit generating the VLP band.

Molina et al. (2008) proposed the conceptual model of the source processes of VLP and LP events recorded at Cotopaxi volcano and located from the waveform inversion (Fig. 15.24). Temporal changes in both the VLP and LP signals were observed, in which a harmonic signature gradually appeared in the LP signal and became dominant over the VLP signal (see Fig. 15.4). Molina et al. (2008) consider that the process began with a magma intrusion beneath Cotopaxi in November 2001, which triggered a swarm of VT earthquakes. The magma intruded into the shallower edifice up to a depth of 2–3 km a.s.l. beneath the northeastern flank of the volcano (Fig. 15.24A). Supersaturated conditions were reached within the intruded magma with a resulting pressure increase through decompression-induced degassing and vesicle growth, while the outer part of the magma solidified due to chilling from the surrounding rock (Fig. 15.24B). As stress due to the pressurization reached the shear strength of the magma, brittle fracture occurred at the top of the magma body, releasing small magma particles with gases. The released

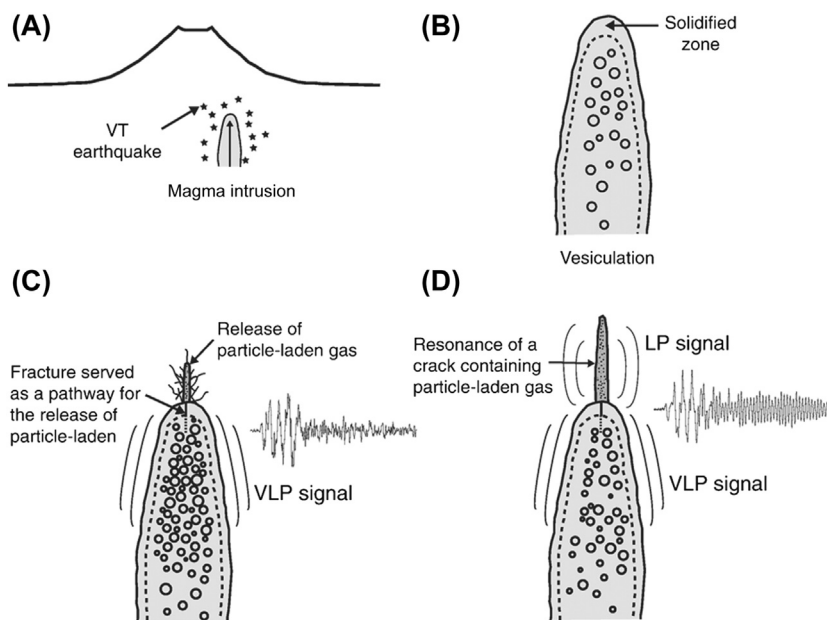


**FIGURE 15.24** Conceptual model for source processes of the very-long-period (VLP) and long-period (LP) events observed at Cotopaxi volcano. See text for details. *From Molina, I., Kumagai, H., García-Aristizábal, A., Nakano, M., Mothes, P., 2008. Source process of very-long-period events accompanying long-period signals at Cotopaxi volcano, Ecuador. Journal of Volcanology and Geothermal Research 176, 119–133.*

particle-laden gas opened a crack in the rock matrix just above the magma system (Fig. 15.24C). The LP events can be interpreted as the resonance of a crack above the magma system, while the VLP events represent the volumetric expansion during the fracturing in the magma and opening of the crack. At the beginning of the VLP and LP events, the LP source region above the magma system consists of a dendritic system of cracks so that the LP waveforms displayed complex nonharmonic oscillations (Fig. 15.24D).

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## Chapter 16

# Seismic Activity Accompanying the Lava Dome Eruptions

Lava domes are mounds of viscous lava and rocks that pile up and accumulate around a volcanic vent. Dome-building eruptions are common, around 6% of eruptions on Earth form lava domes. Lava domes evolve unpredictably, due to non-linear dynamics, caused by crystallization and outgassing of the highly viscous lava in the dome's conduit. These eruptions are among the most hazardous volcanic phenomena because the collapse of domes can result in pyroclastic flows and surges (Melnik and Sparks, 1999; Fink and Anderson, 2000; Calder et al., 2015). The seismicity, characteristic of dome eruptions, is attributed to the excess fluid pressures, which are maintained close to the fracture strength of the dome and wall rock, resulting in fluid movement during the formation of tensile and shear fractures within the dome and upper conduit (Sparks, 1997). In this chapter, I focus on the seismic signals, associated with two end-member styles of lava dome growth: endogenous, in which new lava is added to the interior of an inflating dome, and exogenous, in which new lava accumulates as lobes on the dome surface. We also consider seismic activity during the transition from endogenous to exogenous lava dome growth. The seismic activity, recorded during the episodes of lava dome eruptions at volcanoes Volcán de Colima and Mount St. Helens, is selected.

### 16.1 DYNAMICS OF THE LAVA DOME PROCESS

Lava domes are formed when the erupted lava is too viscous to flow freely away from the vent. Dome diameters vary from a few meters to several km, and their height varies from a few meters to greater than 1 km. The domes may be subdivided into different categories based on morphology, surface texture, and eruptive style (Fink and Anderson, 2000; Hale and Wadge, 2008).

The morphology of lava domes is controlled by magma rheology, crater topography, ascent dynamics, and the mechanism of dome growth. Lava domes comprise core, carapace, and clastic talus components (Fig. 16.1). The carapace represents the region of solidified lava that is attached to the lava core. Talus is composed of loose brittle lava that encloses the massive hot lava dome core (Hale, 2008).

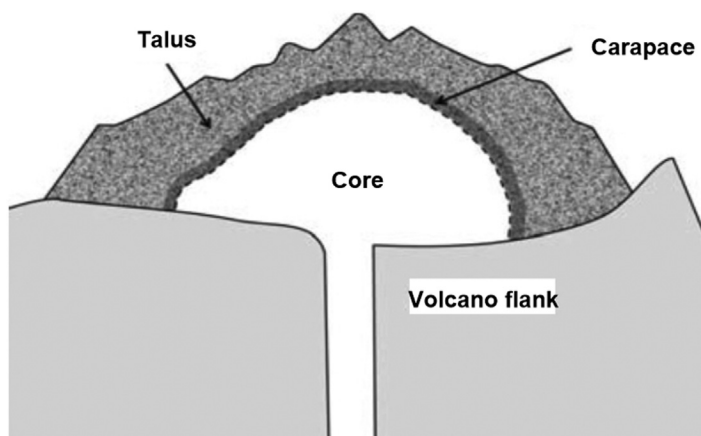


**FIGURE 16.1** Schematic of a simplified endogenous lava dome. *From Hale, A. J., 2008. Lava dome growth and evolution with an independently deformable talus. Geophysical Journal International 174, 391–417*

As was noted earlier, the lava domes can grow endogenously by inflation of a core and/or exogenously with the extrusion of shear bounded lobes and whaleback lobes, and spines at the surface. The shear bounded lobe represents a fault or shear zone-bound unit of lava that extruded from a vent or through the carapace. The whaleback lobe is a type of shear lobe, distinctive because of its smooth arcuate back. The lava spine represents a coherent slab or block of solid lava extruded, usually vertically, from a vent or exogenously through the carapace of a dome. The more core lava that exists within a dome, in both relative and absolute terms, the more explosive energy is available (Hale et al., 2009; Calder et al., 2015).

Two types of lava domes are shown in Fig. 16.2. A photo of the endogenous rhyolite lava dome that was grown in the crater of Chaiten volcano, Chile, in 2010 (<https://earthobservatory.nasa.gov/images/43100/chaiten-volcano-lava-dome-chile>) is presented in Fig. 16.2A. The photo of the exogenous curved-horn andesitic lava spine, growing in 1996 in the crater of Soufriere Hills volcano, Montserrat (Watts et al., 2002), is shown in Fig. 16.2B. The products of endogenous (A, dome) and exogenous (B, spine) growth of lava dome at volcanoes Chaitén, Chile and Soufriere Hills (SHV), Montserrat Island, respectively.

The continuous lava dome-building eruptions consist of a sequence of eruptive episodes separated in time. Lava domes can be formed relatively rapidly, over days to weeks, but eruptive episodes can extend for years to decades alternating between endogenous and exogenous styles of dome constructions. These transitions often coincide with significant changes in the dynamics of magma supply (Watts et al., 2002; Hale and Wadge, 2008; Calder et al., 2015; Fink and Anderson, 2023).

Due to the intermittent buildup of gas pressure, growing domes can often also experience episodes of explosive activity and associated

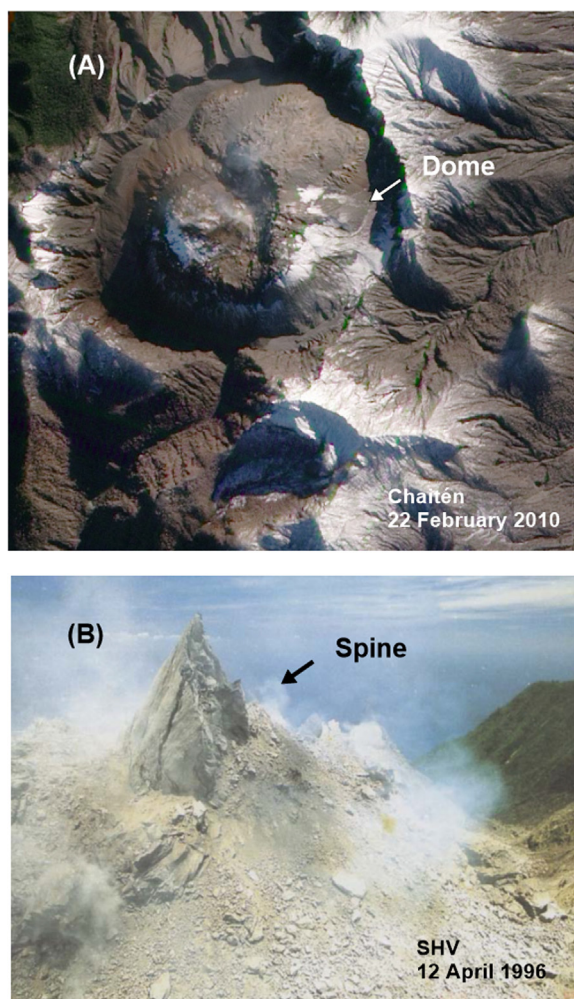


**FIGURE 16.2** The products of endogenous (A, dome) and exogenous (B, spine) growth of lava dome at volcanoes Chaitén, Chile (A) and Soufriere Hills (SHV), Montserrat Island (B), respectively. Photo in A was taken from R. Simmon, NASA (<https://earthobservatory.nasa.gov/images/43100/chaiten-volcano-lava-dome-chile>). Photo in B was taken from Watts, R.B., Herd, R.A., Sparks, R.S.J., Young, S.R., 2002. Growth patterns and emplacement of the andesitic lava dome at Soufrière Hills Volcano, Montserrat. In: Druitt, T.H., Kokelaar, B.P. (Eds.), *The Eruption of Soufrière Hills Volcano, Montserrat, from 1995 to 1999*. Geol Soc Lond Mem. 21, pp. 115–152. <https://doi.org/10.1144/GSL.MEM.2002.021.01.06>.

pyroclastic flows and rockfalls (Wadge et al., 2014; Heap et al., 2019; Zobin and Tellez, 2019).

It is possible to identify the following stages during lava dome eruption: (1) pre-lava stage, or vent clearing; (2) lava dome growth; (3) lava flow formation; (4) lava dome destruction or disintegration (Scott et al., 2008; Wadge et al., 2014).

The first, pre-lava, stage may include a variety of eruptive events that free up space within a volcanic edifice allowing the magma to rise to the surface of the volcano's crater. The start of lava dome growth in the crater is commonly accompanied by diverse explosive volcanic activity. The growth of the lava dome during phase 1 of the 1995–2012 eruption at Soufriere Hills volcano and of the 1991 dome during the 1990–95 eruption at dacitic Unzen volcano, Japan, began after 4 and 6 months of phreatic activity, respectively (Watts et al., 2002; Nakada et al., 1999). The growth of the 1996 lava dome at andesitic Popocatepetl volcano, México, began after 3 weeks of small explosions with ash emissions (Historia . . . , 2012).

The stage of lava dome growth may occur as an endogenous or an exogenous process. In the case of endogenous lava dome growth, the enlargement of a lava dome occurs due to influx of magma into the dome interior. During the exogenous enlargement of a lava dome, vertical lava spines or shear lobes may be extruded through the carapace of a dome (Calder et al., 2015). During any given eruption, dome growth in the crater may occur in different regimes, changing from endogenous to exogenous (Fink et al., 1990). Through the intermittent buildup of gas pressure, growing domes can often experience episodes of explosive activity (Calder et al., 2015; Heap et al., 2019; Zeinalova et al., 2021).

The growth of lava domes commonly transitions to the formation of lava flows when the dome spills out of a crater and elongates downslope (stage 3). The resulting flow fields can extend tens of kilometers long, spread kilometers wide, and reach thicknesses of hundreds of meters, although most are more modest in size. The overwhelming majority have silicic compositions (Kilburn, 2000). Blocky lava flows are characteristic features of andesitic and dacitic lava dome eruptions (Navarro-Ochoa et al., 2002; Harris and Rowland, 2015).

Destruction of the lava dome can occur in the final stages (stage 4) of some dome-forming eruptions. This may be caused by gravitational collapse or explosive decompression of the lava body, disintegration of lava spines, or destruction by Vulcanian explosions.

The processes, associated with different stages of lava dome emplacement, generate seismic signals that give us a repeatable basis for reconstructing the eruptive history. In the following sections, I present the seismic signals, recorded during characteristic endogenous and exogenous dome growth and during the transition from endogenous to exogenous lava dome growth at andesitic Volcán de Colima and dacitic Mount St. Helens volcanoes.

## **16.2 SEISMICITY ACCOMPANYING THE 2016–17 ENDOGENOUS EPISODE OF THE ANDESITIC LAVA DOME ERUPTION AT VOLCÁN DE COLIMA**

The 3860-m-high andesitic stratovolcano of Volcán de Colima is situated at the western end of the Trans-Mexican Volcanic Belt. Volcán de Colima displays a wide spectrum of eruption styles, including small phreatic explosions,

major block-lava effusions, and large explosive events followed by pyroclastic flows and lahars (Bretón González et al., 2002). Its recent persistent dome-building eruption began in November 1998 and was preceded by five swarms of volcano-tectonic (VT) earthquakes recorded from November 1997 (Zobin et al., 2002). These sequences of VT earthquakes potentially constructed a way for juvenile magma to ascend into the shallow magma storage zone which is located at depths between 4 and 12 km beneath the crater (Luhr, 2002; Zobin et al., 2002).

During the 1998–2017 unrest at Volcán de Colima, seven lava-dome building destruction episodes were recorded (Zobin and Tellez, 2019). This section presents the seismic activity that accompanied the endogenous lava dome growth and the subsequent destruction during the seventh, 2016–17, episode.

## 16.2.1 General Description of the 2016–17 Eruption

Fig. 16.3 illustrates the main stages of development of this eruptive episode. Fig. 16.3A shows the empty crater of the volcano in August 2015, as the result of a previous, sixth, episode, which culminated with the destruction of the lava dome. The appearance of a small lava dome in the crater of the volcano, recorded during a helicopter overflight of the crater on February 16, 2016 (Varley, 2016), had heralded Episode 7. This small dome, about 25 m in diameter by 10 m high, was situated within a crater of almost 300 m in diameter (Fig. 16.3B). The lava dome grew for 8 months, filling the crater by the end of September 2016 (Fig. 16.3C) with a volume, estimated to have exceeded 1.2 million cubic meters. On September 30, the lava dome overflowed the crater rim and sent a slow-moving blocky lava flow and incandescent material down the SW flank (Fig. 16.3D). Lava flows, generated between October and December 2016, reached 1700 m in length. Then the destruction of the dome began. A final sequence of large dome-destroying Vulcanian explosions occurred in January–February 2017 (Fig. 16.3E), accompanied by large ash plumes that rose to an altitude up to 10.7 km. The internal crater became about 250 m in diameter and 50–60 m deep (Fig. 16.3F).

## 16.2.2 Seismicity Accompanying Dome Eruption Stages

### 16.2.2.1 System of Monitoring and Seismic Signals

This study is based on seismic data, collected by the seismic network of Colima University (RESCO) from January 2016 to February 2017. Seismograms, obtained from broadband station EZ5, situated 4 km from the crater, were used to determine the properties of eruptive events. Seismograms from short-period station EZV3, situated 6 km from the crater, were used to derive daily counts of the seismic events, produced by





**FIGURE 16.3** Time series of images of the Volcán de Colima lava dome eruption from August 2015 to April 2017. (A) The empty crater of the volcano before the lava dome growth. Photo was taken by C. Navarro on August 12, 2015. (B) The new-born lava dome. Photo was taken by N. Varley on February 16, 2016. (C), the crater of the volcano, filled with new lava dome. (D) The lava flows, appearing after the lava dome filled the crater. Photos in C and D were taken by the Webcams de México on September 28 and September 30, respectively. (E) The largest of explosions, destructing the lava dome. Photo was taken by S. Velasco on January 16, 2017. (F) The empty crater of the volcano after the destruction of the lava dome. Photo was taken on April 28, 2017, by D. Vargas. *Courtesy Colima Volcano Observatory.*

explosions, pyroclastic flows, and incandescent rockfalls. Within the text of this section, the seismic signals and video images of the networks of Colima University are indicated with the UTC time used in monitoring systems.

Explosive activity accompanied the stages of dome growth and destruction. Examples of three types of explosive seismic signals are shown in Fig. 16.4. Two types of seismic signals were generated by Vulcanian explosions. Type 1 seismic signals were generated by explosions, occurring above the lava conduit, and were followed by pyroclastic

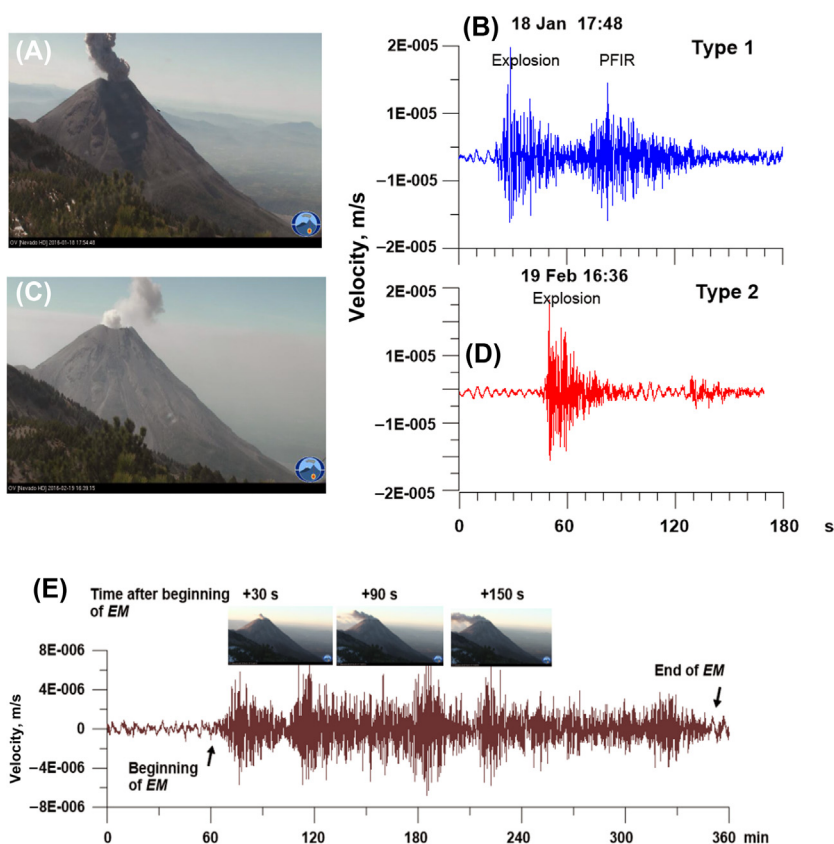


**FIGURE 16.4** Seismic signals generated by explosive events. The seismograms are corrected for sensibility. Two types of broadband unfiltered seismic velocity waveforms, recorded at station EZ5, and video images of corresponding explosive events, recorded at station NEV, are shown in (A)–(D). Video images show the views of explosive plumes, recorded at video station NEV in one minute after their initial appearance. Explosive phase (EXP) and phase of PFIR are indicated. (E) Example of seismic broadband unfiltered velocity signature of EM, recorded at station EZ5 on January 15, 2016. The first member of EM (EM\_1) is indicated with arrows. Video images show the views of explosive plumes, recorded at video station NEV in 30, 90, and 150 s after the beginning of the EM record. Courtesy Colima Volcano Observatory.

flows and incandescent rockfalls (PFIRs) (Fig. 16.4A and B). The PFIRs were emplaced by partial collapse from eruption columns during the type 1 explosion and represented mixtures of rocks and steam-and-ash that accumulated in the crater after the destruction of the previous dome, emplaced in 2015. Type II seismic signals were generated by explosions, occurring within the lava conduit, and were not followed by PFIRs (Fig. 16.4C and D).

A third type of explosive event, explosive multiples (*EMs*), consisted of sequences of explosive-type impulses (Fig. 16.4E, below). Video images, shown in Fig. 16.4 (E, above), demonstrated that *EM* signals were associated with the venting of relatively low-height gas-and-ash plumes. *EM* sequences were not accompanied by PFIRs.

A fourth group of seismic signals was generated by the movement of the PFIRs along the slope of the volcano. The waveform of a seismic signal, produced by rockfall, is shown in Fig. 16.5A. The similar-form signals (Fig. 16.5B) were recorded also during the movement of blocky lava flows, when the collapse of their fronts, and marginal levées-walls, built up on the sides of the lava flow, generate hot avalanches (or lavafalls). The waveforms of lavafalls and of PFIRs differ in their durations and spectral frequencies. Fig. 16.5C gives the comparison of spectral frequencies and durations of the seismic signals of 52 lavafalls, recorded during September–October 2016, and of 62 PFIRs, associated with Vulcanian explosions of type 1, that were recorded in January 2016. The majority of the lavafall signals were characterized by spectral frequencies between 1 and 5 Hz, while the lower spectral frequencies of PFIR signals varied between 0.2 and 2.0 Hz. As is seen also from Fig. 16.5C, the durations of the most of lavafall signals are comparable with relatively short duration signals of PFIRs.

### 16.2.2.2 Processing of the Seismic Signals

Here, we consider the seismic characteristics of explosive events in the context of a two-stage conceptual model of Vulcanian explosions proposed by Zobin et al. (2006) (see Section 11.3.2.1). As is seen in Fig. 11.15B, the seismic signals of explosion events are represented by the two-phase waveforms with an initial low-amplitude phase followed by a high-amplitude high-frequency explosive main phase. According to the two-stage conceptual model, magma enters the conduit, passes the fragmentation zone, and rises to the surface, which generates a subsequent explosion near the surface at the explosion level (Fig. 11.15A). The explosive energy,  $E$ , was estimated from the seismic energy of explosion  $E_{\text{seismic}}$ , determined from Fourier spectra of the main seismic explosive phase, as  $E = E_{\text{seismic}} \times 10^5$  J (Pyle, 2000; Zobin et al., 2006). The errors in the estimation of  $E_{\text{seismic}}$  may reach about  $\pm 10\%$ .

Because the first members of the *EM* multiples, *EM*\_1, were less deformed by seismic noise than subsequent members, they were used for quantification of the events. A comparison of waveforms of *EM*\_1 and of the seismic signals of Vulcanian explosions (Zobin et al., 2023) showed the similarities in their two-phase structures. This comparison allowed us to apply the described methodology for calculating explosive energy, proposed for Vulcanian explosions, to the explosions, that generated the *EM* events. The



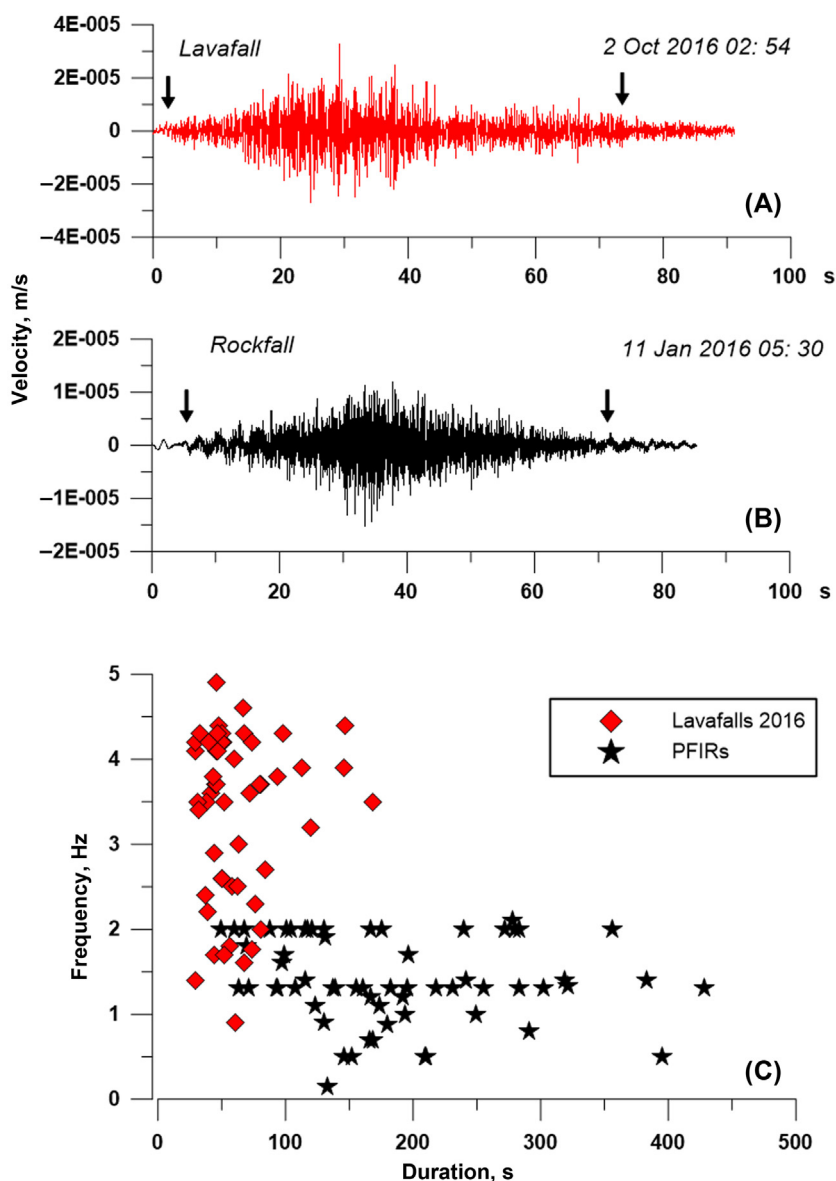


**FIGURE 16.5** Seismic signals of lavafalls and PFIRs. The broadband unfiltered seismic velocity waveforms of lavafall (A) and PFIR (B), recorded at station EZ5, and comparative characteristics of the spectral frequency and duration of the seismic signals of lavafalls, recorded during September–October 2016, and signals of PFIRs, that were associated with the Vulcanian explosions, recorded in January 2016. (C) Arrows in (A) and (B) indicate the duration of signal. *Courtesy Colima Volcano Observatory.*

explosive energy was calculated in each case for the first members of multiples, *EM*<sub>1</sub>.

### 16.2.2.3 Seismicity through the Eruptive Stages of Dome Growth

The seismicity, accompanying the eruption stages of dome growth is analyzed using the temporal variations in the daily number of explosive events, PFIRs, and lavafalls (Fig. 16.6) as well as the temporal variations of energetic parameters of the seismic signals, generated by explosive and effusive processes at different stages of the lava dome growth destruction (Figs. 16.7–16.10).

Stage 1 of lava dome growth (vent clearing) began on January 15, 2016, with a continuous increase of the level of explosive activity from 10 to 25 daily explosive events (Fig. 16.6). This stage 1 continued for about one month, until

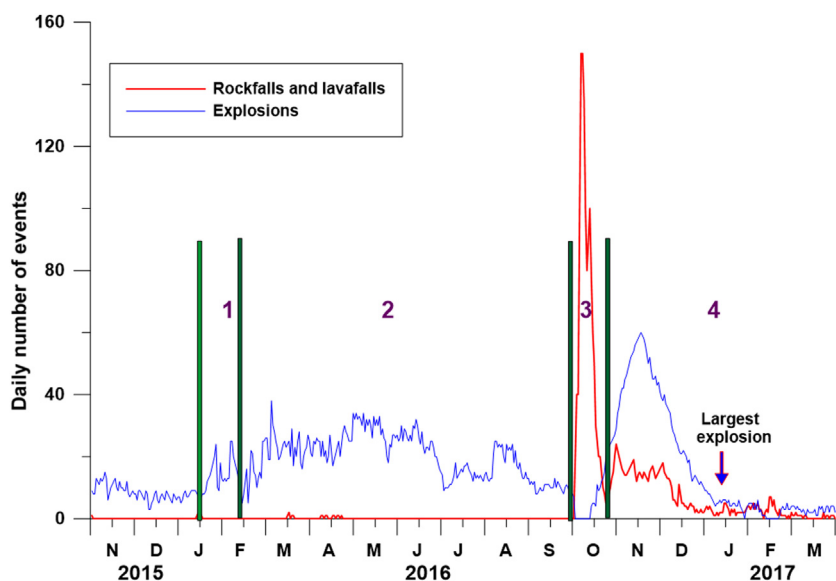


**FIGURE 16.6** Variations in the daily number of seismic events recorded during four stages in eruptive activity of the 2016–17 lava dome eruption at Volcán de Colima. The data were taken using the short-period seismic records at station EZV3, situated 6 km from the crater. Variations in explosive events are shown with the blue line; variations in the number of rockfalls and lavafalls are shown with the red line. The vertical lines separate four stages in eruptive activity, marked with numbers 1 to 4. The moment of the largest explosion, recorded on January 12, 2017 during the stage of lava dome destruction, is shown with a blue arrow. From Zobin, V.M., Brefon, M., Léon, Z., Tellez, A., 2021. Transition from passive degassing to slow growth of a new lava dome as derived from seismic signatures: Volcán de Colima, Mexico, January–February 2016. *Journal of Volcanology Geothermal Research* 420, 107419. <https://doi.org/10.1016/j.jvolgeores.2021.107419>.

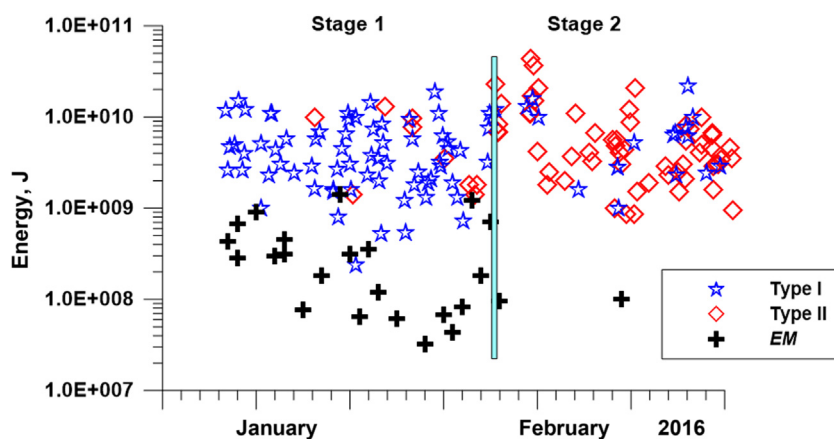


**FIGURE 16.7** Temporal variations of explosive energy of Vulcanian explosions of types 1 and 2 and the first members of EMs. The vertical line indicates the transition from stage 1 of the vent clearing to the initial phase of the lava dome growth at stage 2. *Modified from Zobin, V.M., Arámbula, R., Bretón, M., León, Z., 2023. Explosive multiples preceding the growth of a new lava dome: Volcán de Colima, México, January–February 2016. Journal of Volcanology Geothermal Research 433, 1–8, 107736. <https://doi.org/10.1016/j.jvolgeores.2022.107736>.*

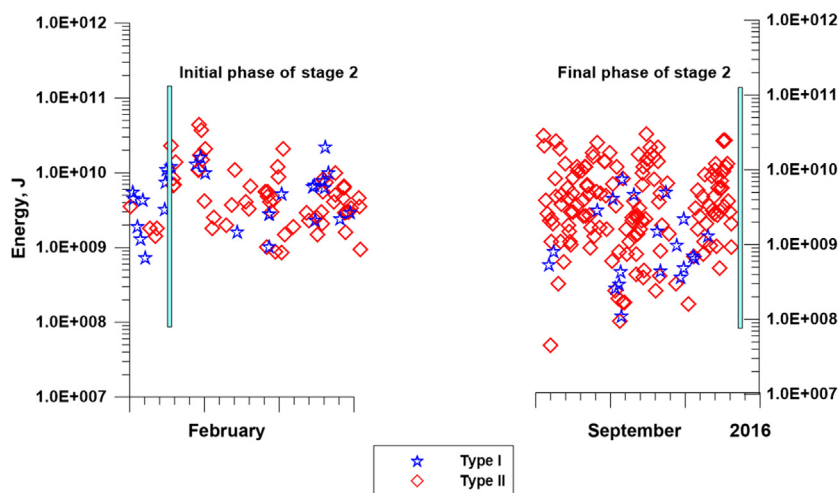


**FIGURE 16.8** Temporal variations of explosive energy of Vulcanian explosions of types 1 and 2 during the initial phase (February) and final phase (September) of lava dome growth at stage 2. The vertical lines separate the events of stage 2 from the events of stage 1 (left) and stage 3 (right). *Courtesy Colima Volcano Observatory.*

the appearance of the new lava dome in the crater in the middle of February, indicating the beginning of stage 2 of the lava dome growth (Fig. 16.3B). Fig. 16.7 illustrates this transition from stage 1 of vent clearing to stage 2 of the

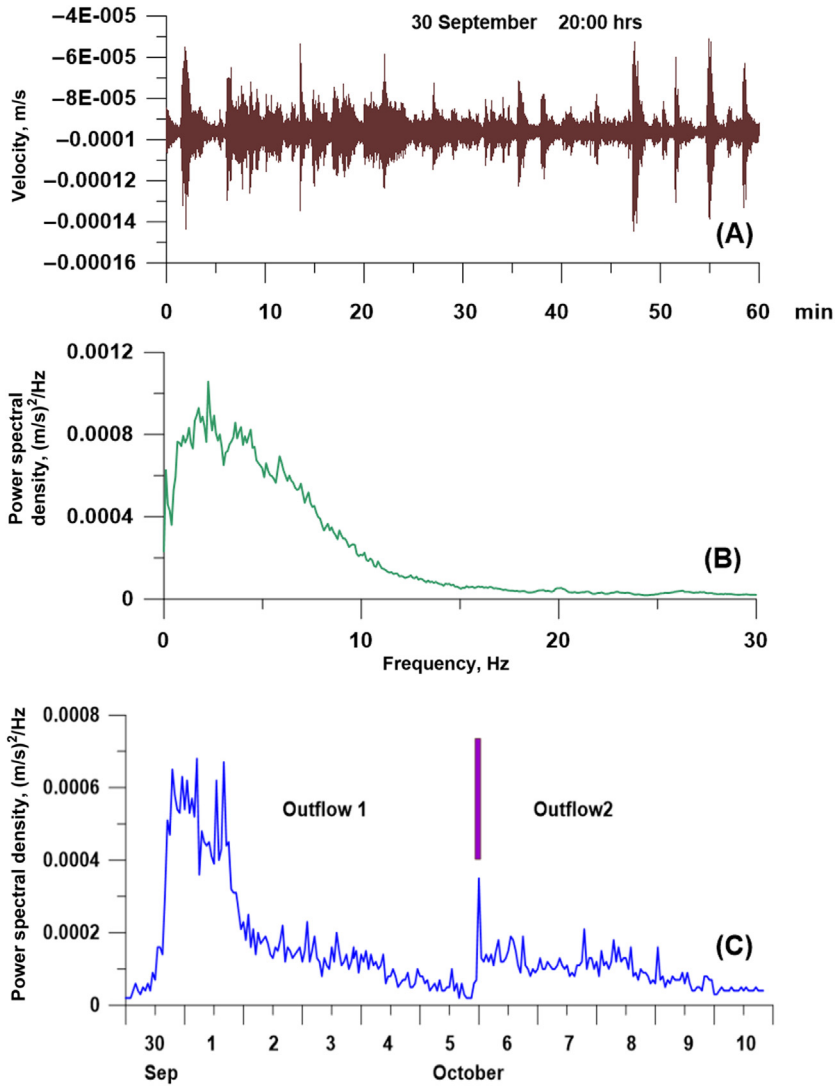


**FIGURE 16.9** Example of the one-hour plot of seismic signals (broadband unfiltered seismic velocity waveform) (A) on September 30, 2016, at 20:00 h, station EZ5, and their power spectral density (PDS) spectrum (B). (C) Dynamics of development of two lava flows occurring from September 30 to October 10, 2016, in terms of the temporal variations of the 1-h PDS spectra amplitudes. Vertical line in C separates two outflows of lava. *Courtesy Colima Volcano Observatory.*

lava dome growth in terms of explosive activity. Total 86 seismic events of type I, 68 events of type II and of 24 EM signatures were considered. Type 1 and EM events dominated during January and the beginning of February

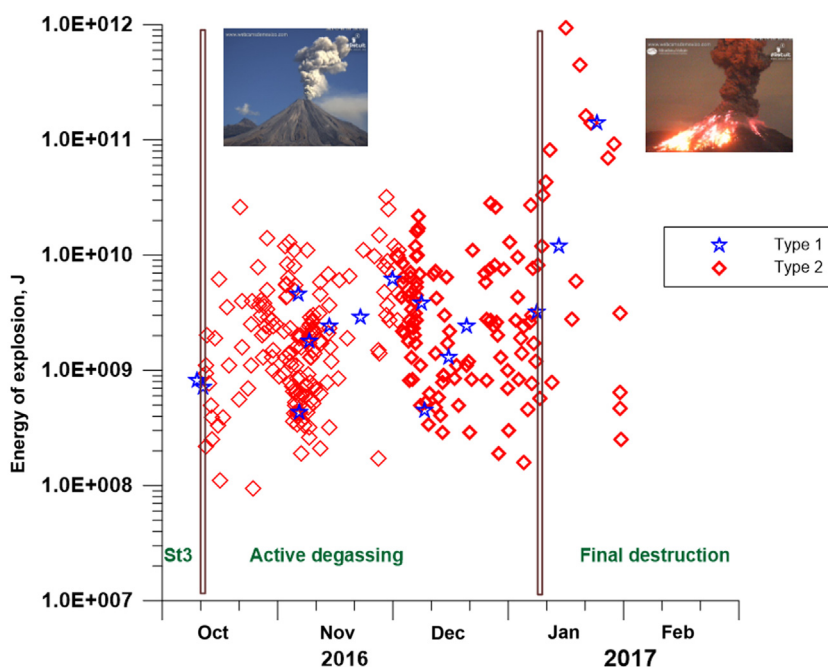


**FIGURE 16.10** Temporal variations of explosive energy of Vulcanian explosions of types 1 and 2 during stage 4. The vertical lines separate the events of stage 4 from the events of stage 3 (St3, left) and indicate two phases of the destruction stage. The characteristic photos of explosive activity during the two phases are shown. *Courtesy Colima Volcano Observatory.*

(stage 1). The type 2 events began to dominate from February 4–6, indicating the beginning of the lava dome growth (stage 2). The explosive energy of type 1 events ranged from  $2.4 \times 10^8$  to  $2 \times 10^{10}$  J, the energy of explosions of type 2 ranged from  $8.6 \times 10^8$  to  $4.4 \times 10^{10}$  J, and that of the first members of the *EM* multiple signals, ranged between  $3.2 \times 10^7$  and  $1.4 \times 10^9$  J (Fig. 16.7B).

Vent clearing during stage 1 was predominated by the explosive events of type 1 explosions and *EMs* (Fig. 16.7), suggesting the following model. The type 1 signals corresponded to the excavation of large masses of rocks in the crater resulting in the appearance of surface faults. Intensive ash-venting along the fault cavities cleared pathways for lava to exit, resulting in the sequences of *EMs*. During stage 2, during the growth of the lava dome, the *EMs* practically disappeared.

The daily number of explosive seismic events in stage 2 grew steadily over the next four months, reaching 30 per day in May (Fig. 16.6). From June to September the daily number of explosions decreased to 10 per day. Stage 2 of the lava dome growth ended at the end of September (Fig. 16.3C).

Fig. 16.8 shows details of the seismic activity of eruptive events during dome growth in 2016. Intermittent buildup of gas pressure led to numerous

episodes of Vulcanian explosive activity. In February 2016, total 18 events of type 1 events with energies, ranging from  $1 \times 10^9$  to  $2 \times 10^{10}$  J and 51 type 2 events with energies from  $7.2 \times 10^8$  to  $4.4 \times 10^{10}$  J. were recorded. In September 2016, there were 20 type 1 events with energies from  $1.1 \times 10^8$  to  $7.4 \times 10^9$  J and 148 type 2 events, ranging from  $4.5 \times 10^7$  to  $3 \times 10^{10}$  J. The increased number of explosive events corresponded to the final phase of the dome building.

Stage 3 (emission of lava flows) began on September 30 (see Fig. 16.3D) and was characterized by a sharp decrease in the number of accompanied explosive events. At the same time, an increase in the number of seismic signals, up to 150 per day, corresponded to the appearance of lavafalls of lava blocks, falling from the flow front and levées during advance of the flow. Intense emission of lava flows continued from September 30 to October 10 (Fig. 16.6). This 11-day interval was selected because later the high-level noise of explosive and PFIR events was not allowed to check the lavafall events. During November and December, only rare emissions of lava flows were recorded.

The study of the development of the lava outflows from the crater down along the volcanic slope was realized in terms of temporal variations of the amplitudes of the power spectral densities (PSD curves), calculated for each one-hour seismic record. Examples of one-hour seismic record at station EZ5 and the corresponding PSD spectrum are shown in Fig. 16.9A and B. The peak amplitudes of one-hour PSD spectra were taken for analysis of temporal variations in the lava flow emplacement. Fig. 16.9C demonstrates the dynamics of the seismic signals, corresponding to two outflows of lava from the crater downslope of the volcano.

The first outflow of lava developed for 5 days, from 13 hours of September 30 to 24 hours of October 5. The level of PSD of the lavafall 1-hour waveform amplitudes began to grow at 17 hours of September 30 and reached its maximum at 19 hours of the same day. This period of large-amplitude waveforms continued for 22 hours. Then the PSD amplitudes gradually decreased.

The next outflow of lava, from 00 hours of October 6 to 24 hours of October 10, began with a single PSD peak. Then the values of PSD amplitudes, corresponding to the following lava emission, continuously decreased.

As is seen, the lava emission during the two outflows had a different character. The first outflow began with a 22-hour period of large-amplitude PSD of 1-hour signals. Outflow 2, which developed during the same 5-day time interval as outflow 1, only exhibited one large-amplitude PSD signal, recorded during its first hour of emission.

The gradual destruction of the lava dome (stage 4) began in November with a sharp increase in the number of explosive events (up to 50 per day) against decreasing in the number of lavafalls waveforms (Fig. 16.6). Active degassing continued through the end of December. In January–February

2017, the frequency of explosive events decreased to 3–5 daily. During this time, the lava dome was destroyed by large Vulcanian explosions (Fig. 16.3E and F).

Fig. 16.10 shows how explosive energy varied during stage 4. Active degassing continued from October 10, 2016, to January 8, 2017. Type 2 explosions dominated in number and energy. The maximum energy of type 2 explosions during this phase reached  $2.8 \times 10^{10}$  J. A final phase of lava dome destruction was caused by a sequence of five large Vulcanian explosions, which began on January 8. The largest of explosions occurred on January 12, having an energy of  $9.4 \times 10^{11}$  J and emptying the crater (see Fig. 16.3F).

### 16.3 SEISMICITY ACCOMPANYING THE 2004–08 EXOGENOUS EPISODE OF THE DACITIC LAVA DOME ERUPTION AT MOUNT ST. HELENS

Mount St. Helens (MSH) is a 2950-m-high dacitic stratovolcano within the Cascade Range of the western state of Washington, USA. It is the only volcano in the lower 48 states to have had significant eruptions in the past 50 years, with dome-forming activity in 1980–86 and 2004–08. Following the growth of a large bulge on the north side of the mountain, eruptive activity began with a steam-blast eruption from the summit crater on March 27, 1980 (see Section 7.1.1). It was followed by a sequence of small explosive eruptions up to the climactic Plinian eruption of May 18. A new crater, with an area of  $1.5 \times 3.0 \text{ km}^2$  and a depth of about 500 m, was formed in the northern flank of the volcano. Lava domes began to grow in the crater in June 1980. During 1980–86, 19 eruptive episodes built a new 267-m-high composite lava dome. Episodes typically began with a period of endogenous distension, followed by extrusion of an exogenous lobe. When the eruption ceased in 1986, a 1-km diameter, 350-m-high lava dome (Fig. 16.11A) with a volume of  $74 \times 10^6 \text{ m}^3$  had been constructed (Fink et al., 1990; Moore et al., 1981). After an 18-year hiatus, renewed activity began on October 11, 2004, resulting in the continuous growth of a complex lava dome of uniform chemical composition accompanied by persistent but low levels of seismicity, rare explosions, low gas emissions, and frequent rockfalls (Scott et al., 2008).

#### 16.3.1 General Description of the 2004–08 Eruption

The reawakening of Mount St. Helens was heralded by a swarm of shallow (depth <2 km) volcano-tectonic earthquakes below the summit of the volcano beginning on September 23, 2004 (Moran et al., 2008). This unrest initiated the steady, continuous extrusion of gas-poor dacitic magma through glacier ice, producing a lava dome located between the 1980 and 1986 dome and the south crater wall (Fig. 16.11B). The newly formed dome consisted of a series of recumbent, smoothly surfaced spines, or “whalebacks” (Fig. 16.11C and D).



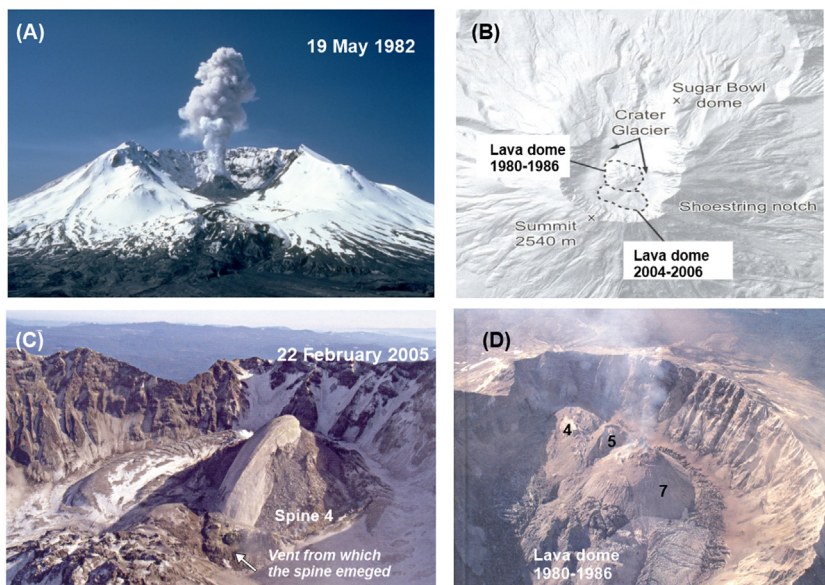


**FIGURE 16.11** Time series of images of the 2004–06 Mount St Helens lava dome eruption. (A) The lava dome of 1980–82. (B), Comparative position of the lava domes of 1980–82 and 2004–06. (C), A view of spine 4. (D), Oblique aerial photography looking SW into the crater of Mt St Helens and showing the system of spines 4, 5, 6, and 7. *From Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), 2008. A Volcano Rekindled; the Renewed Eruption of Mount St. Helens, 2004–2006: U.S. Geological Survey Professional Paper 1750, 856 p. (A) Taken by L. Topinka, USGS. (B) Taken from Major et al., 2008. (C) Taken by S.P. Schilling, USGS. (D) Photographs taken by J.S. Pallister, USGS.*

During 2004–08, a sequence of seven solidified lava spines developed in the crater of the volcano (Iverson et al., 2006; Scott et al., 2008; Vallance et al., 2008; Major et al., 2009).

The spines repeatedly formed and disintegrated. Two minor spines (1 and 2) emerged in October 2004. Spines 3 and 4 subsequently grew and crumbled between late October 2004 and April 2005. Spine 5 emerged, grew, and crumbled between April and August 2005. Growth of spine 6 followed, but it was rapidly subsumed by the emergence and enlargement of one other major and several smaller spines, which then combined with endogenous growth to form a broad mass collectively called spine 7, which emerged in October 2005 and grew continuously until January 2008. The eruption ended in 2008, having built  $92 \times 10^6 \text{ m}^3$  lava domes (Major et al., 2009; Vallance et al., 2008).

### 16.3.2 Seismicity Accompanying the Eruption Stages

The 2004–08 extrusion was driven by a nearly constant influx of magma at depth and was resisted by a plug of solidified magma that slipped



incrementally and seismogenically against the wall rock forming the upper parts of the magma conduit. By the time magma reached 1 km below the surface, it was nearly completely degassed and crystallized (Sherrod et al., 2008; Calder et al., 2015). This led to a practical absence of outgassing seismicity of explosive events. The 2-km-deep seismicity, recorded during the eruption, likely represented the development of a fracture system in the base of the plug (Moran et al., 2008).

### 16.3.2.1 System of Monitoring and Seismic Signals

The seismic events, occurring during the 2004–08 eruption at Mount St. Helens, were recorded by the short-period vertical-component seismometers of the Pacific Northwest Seismic Network (PNSN). The PNSN presented the catalog of located earthquakes and, when earthquakes had become too numerous, the real-time seismic amplitude measurement (RSAM). Magnitude  $M_d$  of seismic events is based on the duration of the time decay of the amplitude of the seismogram and is used in USGS for very small events (<https://www.usgs.gov/programs/earthquake-hazards/magnitude-types>).

The seismic events were represented mainly by volcano-tectonic, hybrid, and low-frequency earthquakes. Their waveforms are shown in Fig. 16.12.

### 16.3.2.2 Through the Stages of the Eruption

Fig. 16.13 shows the two stages of the lava dome-building eruption as derived from the temporal variations of the RSAM. Two stages are indicated: stage 1 (September 23–October 10) of pre-lava vent clearing and stages 2–4 (October 10, 2004, to December 2005) of lava-spine growth (stage 2), followed by their destabilization or disintegration (stage 4). Stage 3 of the emission of lava flows was absent.

Fig. 16.14 shows variations in the number of seismic events and their maximum daily magnitudes during the vent clearing, stage 1. The seismic activity developed into three phases, reflecting the movement of dacitic lava to the surface. During phase 1 (September 23–27), a small swarm of shallow (<2 km) volcano-tectonic earthquakes were recorded, with a maximum magnitude of  $M_d = 2.7$ . This initial seismic activity corresponded to the appearance of a series of cracks in the glacier, situated immediately south of the 1980–86 lava dome (Moran et al., 2008; Scott et al., 2008; PNSN catalog).

In phase 2 (September 28 to October 6), a gradual transition from volcano-tectonic to hybrid and low-frequency (LF) seismic events occurred. The number of events sharply increased up to 120 on September 29. The maximum magnitude reached  $M_d$  3.8 on October 2. Two energetic non-eruptive tremor episodes were recorded on October 2 and 3. Phase 2 culminated with four phreatic eruptions between October 1 and October 5. A small explosion on October 1 formed a vent through the

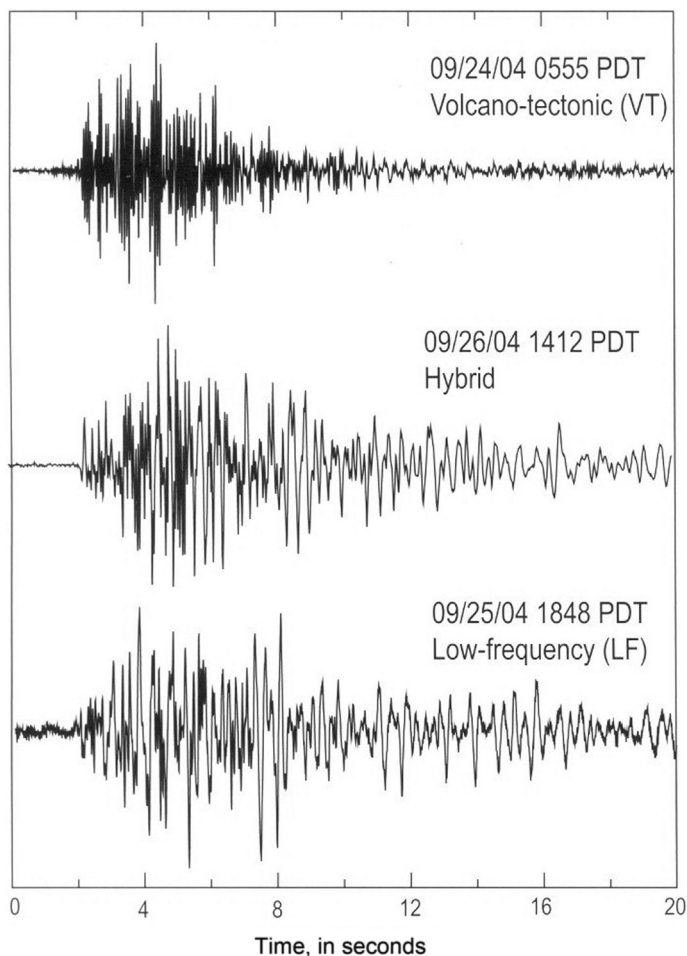


**FIGURE 16.12** Waveforms for typical volcano-tectonic (VT), hybrid, and low-frequency (LF) events recorded at short-period station HSR of Pacific Northwest Seismic Network (PNSN), situated 3 km west of the vent. From Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Prof. Paper 1750, Washington, pp. 27–60.

~150-m-thick glacier that had grown in the southern part of Mount St. Helens' post-1986 crater. The last explosion was the largest, generating an ash plume that reached about 1 km above the crater rim and rated a Volcanic Explosivity Index (VEI) of 2. Following this last explosion of the series, phase 2 seismicity declined significantly (Moran et al., 2008; Scott et al., 2008; PNSN catalog).

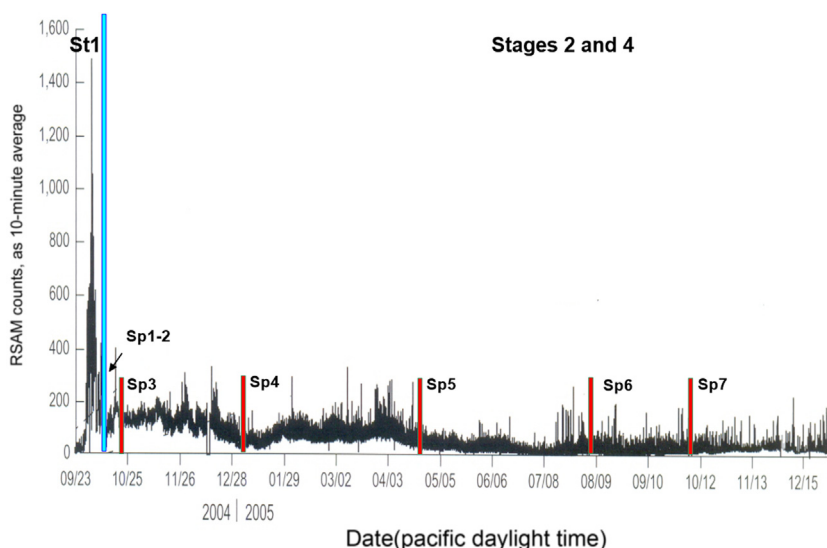


**FIGURE 16.13** The stages of lava dome-building eruption at Mount St Helens as derived from the temporal variations of the 10-min real-time seismic amplitude measurement (RSAM) values at short-period station SHW, 3 km west of vent. *Sp1–7* indicate spine formation. *St1*, stage 1. Tall blue vertical line separates stages 1 and 2–4. Low vertical lines mark the moments of appearance of seven spines. Modified from Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W. A., Wright, A.K., Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St.Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Prof. Paper 1750, Washington, pp. 27–60.

Over the next few days (phase 3, from October 7 to October 10), seismicity was dominated by several event families consisting of low-magnitude hybrid and LF earthquakes. Seismicity declined significantly following the October 7–8 earthquake sequence. On October 10, a fixed camera on the Sugar Bowl dome, at the northeast mouth of the crater, began transmitting images that tracked the growth of part of the welt, formed on the southeast margin of the previous lava dome. Later analysis of digital elevation models showed that by October 11 the deformed area had a volume of about  $10 \times 10^6 \text{ m}^3$  (Scott et al., 2008). An area on the northwest part of the welt had reached about  $300^\circ\text{C}$  by October 10, suggesting that the crater floor was being heated. As was noted by Moran et al. (2008), the appearance of seismic event families, taken together with the overall decline in seismic levels, the change in event character to dominantly LF and hybrid events, the relatively low gas levels, and the observed deformation of the crater floor indicated that these LF events likely represented the extrusion of a new dome.

With beginning of the stage 2 on October 11, 2004, the new dome developed as a series of recumbent, smoothly surfaced whalebacks, that extruded

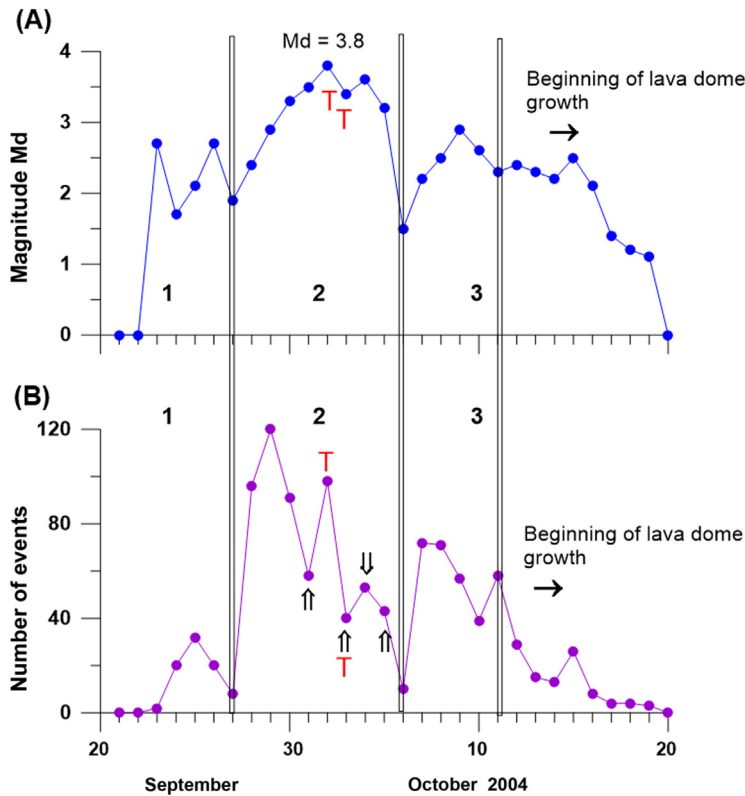


**FIGURE 16.14** Variations of the maximum daily magnitudes of seismic events (A) and their daily number (B) during stage 1 of vent clearing. Three phases of this stage are separated with vertical lines and marked with 1–3. Episodes of phreatic explosions are indicated with arrows. Episodes of seismic tremor are indicated with T. Data from the PNSN catalog.

to lengths of almost 500 m (see Fig. 16.11C). Each of the seven spines was mantled by ash-gouge, bordered by a zone of highly localized deformation that included tensile and shear fractures, cataclasite, and occasionally pseudo-tachylite. Fig. 16.15 illustrates the seismic activity of the initial extrusion of the lava spines. The appearance of fin-shaped rock spine 1 about 30 m high and 60 m long on October 11 was accompanied by small ( $M_d < 1$ ) hybrid events, occurring at a rate of 1–2 per minute, as well as slightly larger events with longer between-event intervals. These multiples of hybrid earthquakes faded before the extrusion of spine 2 on October 16.

With the extrusion of spine 2, there were began to appear the regularly spaced, small earthquakes. These seismic events had occurred with such remarkably constant intervals that they were named by seismologists from USGS (Moran et al., 2008) as “drumbeat” earthquakes (Fig. 16.16A). The drumbeats consisted of repetitive low-frequency and hybrid earthquakes, like

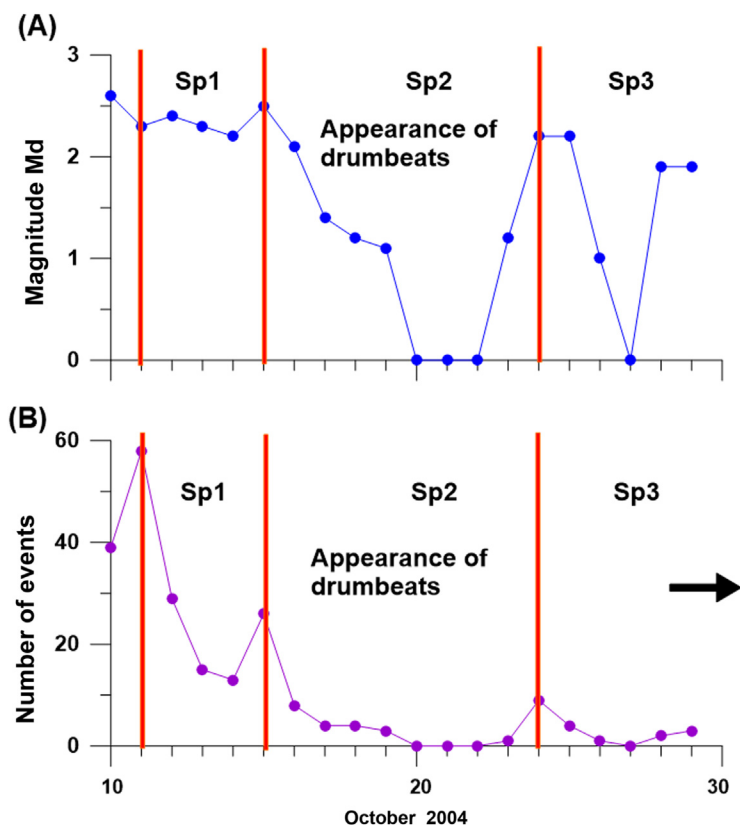


**FIGURE 16.15** Variations of the maximum daily magnitudes of seismic events (A) and their daily number (B) during period of appearance of the first three lava spines. Periods of growth of these lava domes are separated by vertical lines and marked with Sp1–Sp3. Horizontal arrow indicates continuation of the growth of spine 3. From Vallance, J.W., Schneider, D.J., Schilling, S.P., 2008. Growth of the 2004–2006 lava-dome complex at Mount St. Helens, Washington. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The First Year of Renewed Eruptions at Mount St. Helens, 2004–2006*. U.S. Geol. Surv. Prof. Pap., 1750, pp. 169–208. Data from PNSN catalog.

waveforms, shown in Fig. 16.12, but significantly lower in amplitudes (Fig. 16.16B). The hybrid drumbeats were relatively rare and occurred in temporal clusters. Such clusters corresponded to transitional periods when one spine became inactive, and a new spine began to grow (Moran et al., 2008). Sherrod et al. (2008) proposed that stick-slip motion was responsible for the seismicity of drumbeats, linking the events to slip events along the conduit margin, although their precise source mechanism remains elusive.

Drumbeats were occasionally accompanied by larger, or “big” ( $M_d > 2.0$ ,  $M_{\max} = 3.4$ ), shallow earthquakes (Fig. 16.16A). These earthquakes were

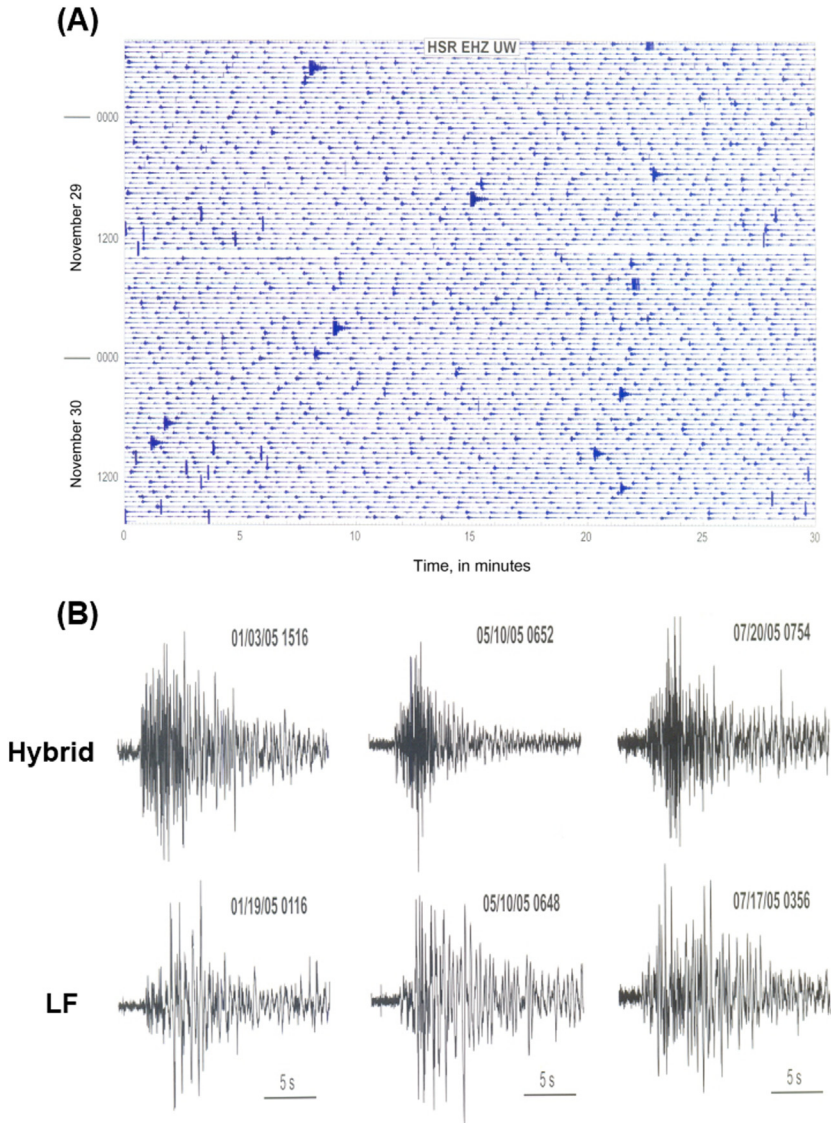


**FIGURE 16.16** (A) Seismic record from station HSR, situated 3 km west of the vent of the new dome, for the 48 h between 1700 PST on November 28 and 1700 PST on November 30, 2004, showing regular drumbeats and 10 “big” earthquakes occurring along them. (B) Examples of waveforms of hybrid (top) and LF drumbeats (bottom) as recorded at station HSR in January, May, and July 2005. From Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., and Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Prof. Paper 1750, Washington, pp. 27–60.

notable because they were much larger than the regular drumbeat events. No change in drumbeat spacing or size occurred before or after the larger earthquakes, indicating that source processes were separate and distinct from one another. This type of seismicity, represented by drumbeats and big earthquakes, accompanied the total process of the lava-spine growth during 2004–05.

Growth and disintegration of spines at Mount St. Helens progressed steadily from October 11, 2004, apparently without pause, and continued throughout 2005. Processes of spine growth and disintegration were not marked with any special change in the seismic activity. Fig. 16.13 illustrates the variations in drumbeat's activity during the growth of all seven spines. The dome grew through the emergence, growth, and collapse of a sequence of solidified lava spines, accompanied by seismic activity at all stages. It is not possible to see any regularity in their development. Disintegration of the spines was also unaccompanied by any regularity in the seismic process. Fig. 16.17 shows the process, leading to the disintegration of spine 5 at the end of its growth on April 19, 2005, and in the following six months. It was significantly destroyed but no significant changes in seismic activity were observed.

In January 2008, the dome stopped growing, leaving three disintegrated spines numbers 4, 5, and 7 (see Fig. 16.11D) (Major et al., 2009; Dzurisin et al., 2015).

## 16.4 SEISMICITY ACCOMPANYING OF THE TRANSITION FROM ENDOGENOUS TO EXOGENOUS ANDESITIC LAVA DOME GROWTH: VOLCÁN DE COLIMA, 2010–11

Dome growth in craters may alternate between endogenous to exogenous styles. The lava dome-building eruption of 1998–2017 at Volcán de Colima, which was generally characterized by seven episodes of endogenous lava dome construction and destruction (one of which was discussed in the first section of this chapter), had also undergone two cases of exogenous lava enlargement. The first of these occurred in October–November 2001. A light-colored spine about 60 m tall and 40 m wide at its base emerged on the N side of the crater. The second occurred during the 2007–2011 lava dome growth and was associated with the emplacement of a lobe, extruded on the dome's margin within the western sector of the volcanic crater from February 2010 to March 2011 (Zobin et al., 2017). This section focuses on this second case.

### 16.4.1 General Description of the 2007–11 Eruption

The lava dome growth during the 2007–11 episode at Volcán de Colima developed from January 2007 to June 2011 (Zobin and Tellez, 2019). The appearance of a new lava dome was first observed at the beginning of February 2007 (Fig. 16.18A). It grew slowly and had filled the crater in two



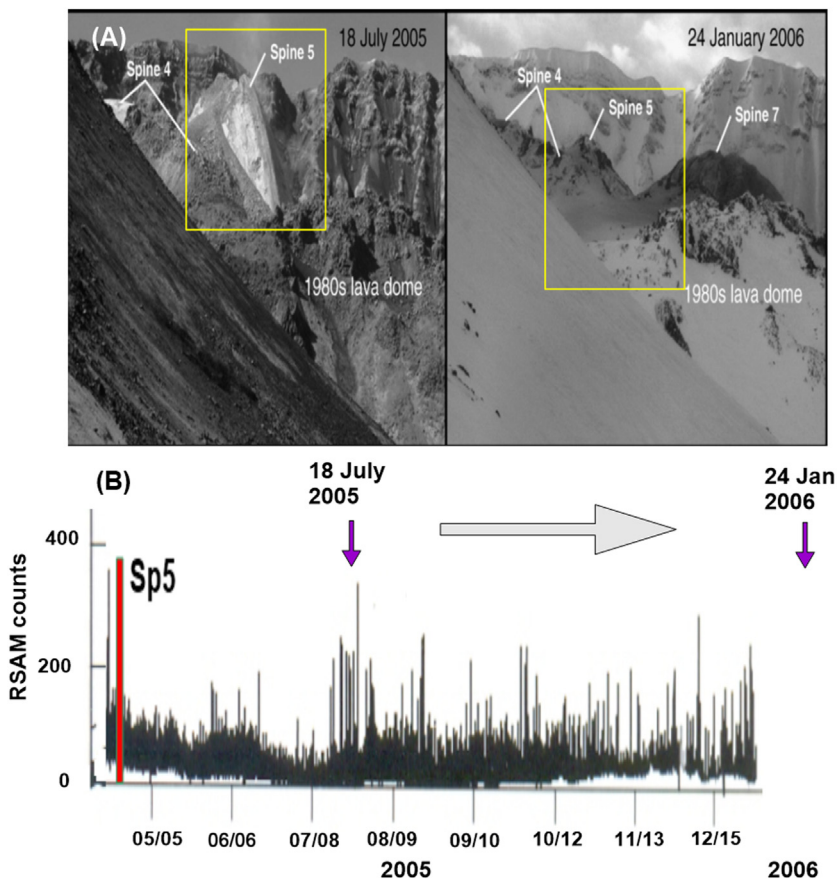


**FIGURE 16.17** Illustration of the process of disintegration of spine 5. (A) shows (with arrows) photos of this spine taken on the date of its complete growth on July 18, 2005 (left) and its disintegration six months later on January 24, 2006 (right). (B) Temporal variations of RSAM counts during this period are shown. For A. From Major, J.J., Dzurisin, D., Schilling, S.P., Poland, M. P., 2009. Monitoring lava dome growth during the 2004–2008 Mount St. Helens, Washington, eruption using oblique terrestrial photography. *Earth Planet Science Letters* 286, 243–254. <https://doi.org/10.1016/j.epsl.2009.06.034>. For B. From Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J. (2008a). Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Prof. Paper 1750, Washington, pp. 27–60.

years (Fig. 16.18B and C). In fall 2009, the new dome reached the western rim of the crater (Fig. 16.18D). The growth rate during 2009 and through March 2010 was low, about  $\sim 0.02 \text{ m}^3/\text{s}$ . The initial endogenous growth of the lava dome was complicated in February 2010 by an exogenous growth of a small lobe of lava within the western sector of the crater of the volcano



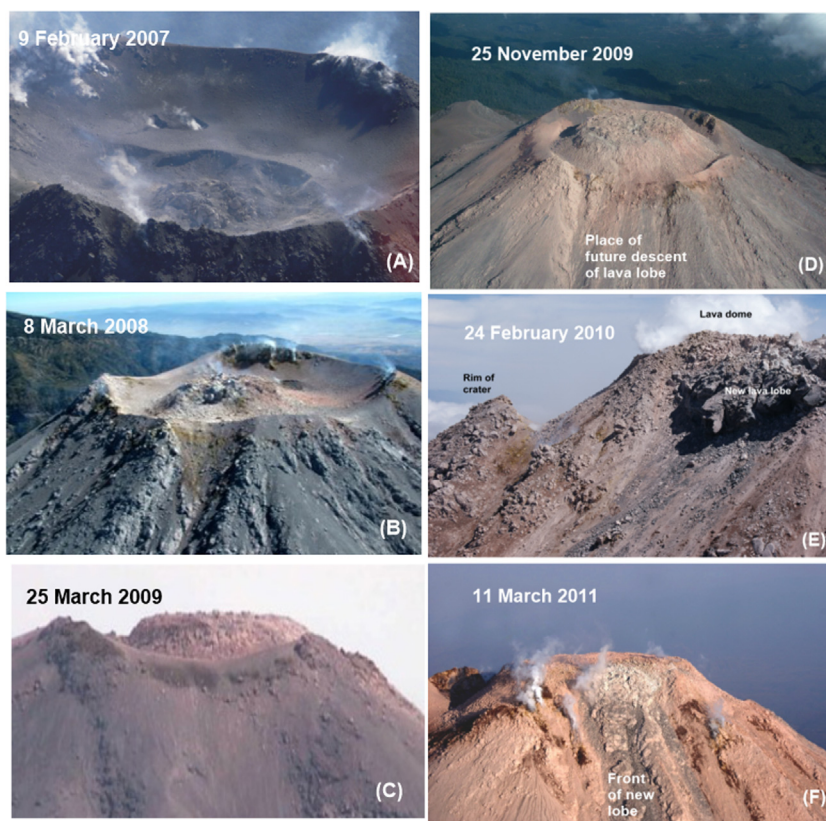


**FIGURE 16.18** Time series of images of the Volcán de Colima lava dome eruption from February 2007 to February 2011. (A), the new-born lava dome. (B), the crater of the volcano with a newly growing lava dome. (C), the crater of the volcano filled with the new lava dome. (D–F), Aerial photos of Colima’s crater zone taken (top to bottom) on November 25, 2009, February 24, 2010, and March 11, 2011, illustrating the appearance of the lava lobe. All three are views from the volcano’s W side. The photos show (D) the endogenous lava dome before the initiation of a new exogenous lava lobe (November 25, 2009), (E) a new lava lobe appearing against the background of the endogenous lava dome (February 24, 2010), and (F) lava lobe development to March 11, 2011. *Courtesy Colima Volcano Observatory. (A) Photo taken by N. Varley on February 9, 2007. (B) Photo taken one year after the beginning of lava dome growth. From BGVN (2008). (C) Photo taken two years after the start of lava dome growth. From BGVN (2009). (F) Photos taken by C. Navarro.*

(Fig. 16.18E). The lobe, which first perched on the dome’s margin, subsequently traversed across the buried crater rim and began to descend the outer flank (Fig. 16.18F). The rate of growth of the lobe was  $\sim 0.1 \text{ m}^3/\text{s}$ , or five times greater than the rate of the dome extrusion. The active lobe’s width reached  $\sim 10\text{--}15 \text{ m}$ ; its length was  $\sim 50\text{--}60 \text{ m}$ . The simultaneous growth of the lava dome and lava lobe stopped completely in

June 2011 (Zobin et al., 2017). The lava dome was destroyed in 2013 with Vulcanian explosions at the very beginning of the next eruption episode (Zobin et al., 2015).

## 16.4.2 Seismicity Accompanying the Eruption Process

### 16.4.2.1 System of Monitoring

The study is based on seismic data collected by the seismic network of Colima University (RESCO) from January 2007 to February 2011. The signals were recorded by two short-period seismic stations, EZV4 and EZV5, situated at 1.7 km and at 4 km from the vent, respectively. The seismograms, obtained from station EZV5, were used for processing the properties of the eruption events. Seismograms, obtained from station EZV4, were used for daily count of the seismic events, produced by explosions, pyroclastic flows, and incandescent rockfalls.

### 16.4.2.2 Development of Eruptive Process Derived from Seismic Monitoring

Fig. 16.19 illustrates the development of eruption processes based on the daily count of seismic events (Zobin et al., 2017). Permanent explosive activity characterized the entire endogenous phase of growth. During 2007 and 2008, when the dome was growing within the crater, small background explosions were recorded daily. Low-level explosive activity continued in 2009.

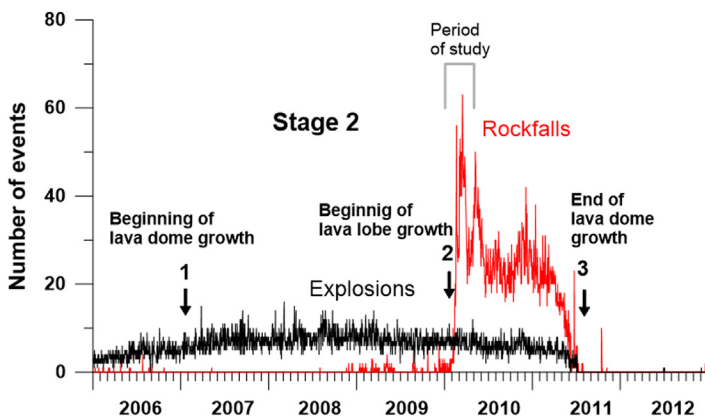


**FIGURE 16.19** Characteristics of eruptive activity during the 2006–2011 eruption at Volcán de Colima (stage 2), based on the counts of daily seismic signals, associated with small explosions (black) and rockfalls (red). Arrows 1, 2, and 3 indicate the principal moments in the growth of the lava dome. The period of study discussed in this section is shown. From Zobin, V. M., Tellez, A., Aguilar, J.E., Cruz, K.G., 2017. Seismic signals of rockfalls as indicators of the origin of lava fragments emplaced during the 2010 endogenous and exogenous growth in the crater of Volcán de Colima, México. *Bulletin of Volcanology* 79, 47, 1–9. <https://doi.org/10.1007/s00445-017-1130-z>.

When the lava dome filled the crater in early 2009 (Fig. 16.18C), low-level rockfall activity at the margins of the dome top began to be recorded (Fig. 16.19). A sharp increase in the number of rockfalls in February 2010 coincided with the beginning of the lava lobe growth and its descent down the outer flank of the volcanic cone. Rockfall activity stayed at a high level during the rest of the lava lobe development.

Pyroclastic flows and rockfalls generate similar emergent, cigar-shaped waxing and waning seismic signals with frequencies of 1–10 Hz, which differ only in their durations and/or amplitudes (see Chapter 13). The rockfalls, that occur during periods of low activity of lava dome eruptions (slow dome growth), are considered as passively generated. Actively generated rockfalls, in turn, are associated with active shear lobes (Calder et al., 2002). The transition from endogenous to exogenous styles of dome growth at Volcán de Colima in 2010–11 represents a case of interaction of passively and actively generated rockfalls.

The spectral content of seismic signals, generated by both types of rockfalls, was chosen as a criterion for the identification of the differences between waveforms, characteristic of the endogenous and exogenous origin of rockfalls (Zobin et al., 2017). The period of January to April 2010 (Fig. 16.20), which

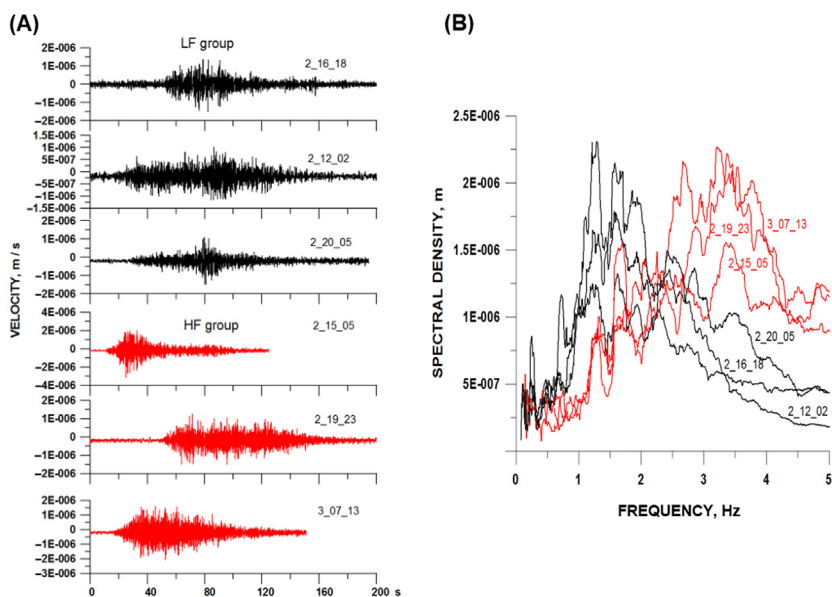


**FIGURE 16.20** Examples of the seismic signals (velocity, vertical component, SP station EZV5) of rockfalls of LF (black) and HF (red) groups of the February–April 2010 events (A) and their corresponding Fourier spectra (B). The indexes near the signals and spectra are shown as month\_day\_hour. From Zobin, V.M., Tellez, A., Aguilar, J.E., Cruz, K.G., 2017. Seismic signals of rockfalls as indicators of the origin of lava fragments emplaced during the 2010 endogenous and exogenous growth in the crater of Volcán de Colima, México. *Bulletin of Volcanology* 79, 47, 1–9. <https://doi.org/10.1007/s00445-017-1130-z>.

includes a period of continuous lava dome growth in January and three months of active enlargement of the lava lobe, accompanied by continuous lava dome growth, was taken into account. This interval reflects the transition from endogenous to exogenous andesitic lava dome styles of growth and was characterized by a predominance of rockfall activity.

#### 16.4.2.3 *Seismic Signals of Rockfalls as Indicators of the Origin of Lava Fragments, Emplaced during Endogenous and Exogenous Dome Growth*

A total of 410 seismic signals of rockfalls were selected for study, 61 during endogenous lava dome growth in January 2010 and 339 during the endogenous lava dome growth in February–April 2010.

Spectral analysis of 339 rockfall signals from February to April 2010 allowed to divide them into two groups: LF with spectral peak frequencies ranging between 1.0 and 2.0 Hz, and high-frequency (HF) with spectral peak frequencies ranging between 3.0 and 4.0 Hz. The characteristic seismic waveforms of rockfalls and corresponding Fourier spectra of the signals of the two groups are presented in Fig. 16.20A and B.

To check the statistical difference between samples of the LF group with a mean peak frequency of  $1.76 \pm 0.44$  Hz and the HF group with a mean peak frequency of  $3.13 \pm 0.63$  Hz, the Kolmogorov-Smirnov (KS) test (<http://www.physics.csbsju.edu/stats/KS-test.html>) was applied considering as a null hypothesis that these two databases,  $n_1 = \text{LF}$  and  $n_2 = \text{HF}$ , belong to the same sample. The maximum difference between two cumulative distributions,  $D_{n_1, n_2}$ , was calculated to be 0.8394. To reject the null hypothesis at the significance level  $\alpha$ ,  $D_{n_1, n_2}$  must be greater than the critical value equal to  $k_{1-\alpha}\sqrt{(n_1 + n_2)/n_1n_2}$ , where  $k_{1-\alpha}$  is the quantile of the KS distribution at the significance level  $\alpha$  (Muller et al., 1979). The significance of  $D$  was checked at the level of  $\alpha = 0.01$  ( $k_{1-\alpha} = 1.628$ ), yielding the critical value  $D_{\text{crit}}$  of 0.1184, which is smaller than the observed  $D_{n_1, n_2}$ . It allows us to consider that the databases of LF and HF seismic signals of rockfalls belong to different groups of signals and may be considered as having different origins.

The distributions of peak frequencies of 61 signals, recorded in January, and 339 signals, recorded during February–April, show that if the February–April events belonged to both groups, including LF and HF events, with most events in the HF group, then the January events are represented mainly by LF events (Fig. 16.21A). Application of the KS-test (Fig. 16.21B) supports that the majority of January rockfall signals with a mean peak frequency of  $1.50 \pm 0.43$  Hz belong to the LF group at the 0.01 significance level. Therefore the LF signals may be associated with passively generated rockfalls, emplaced during the gravitational collapse of the marginal parts of a growing endogenous lava dome during the January and February–April periods. In turn, the HF signals of rockfalls were likely

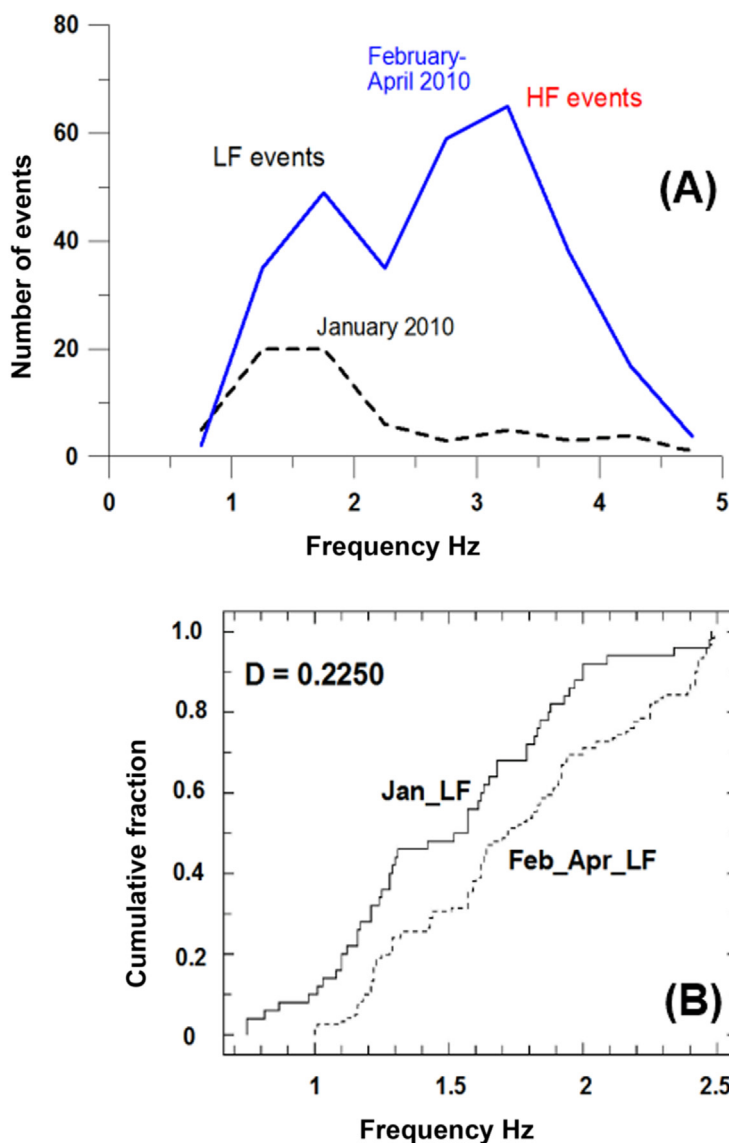


**FIGURE 16.21** Distributions of the Fourier spectral peak frequencies of the seismic signals of rockfalls, recorded in January (*dashed line*) and February–April (*continuous line*) 2010 at short-period (SP) station EZV5 (A), and application of KS\_test to samples of LF seismic signals recorded during January (Jan\_LF) and February–April (Feb\_Apr\_LF) 2010 (B). From Zobin, V. M., Tellez, A., Aguilar, J.E., Cruz, K.G., 2017. Seismic signals of rockfalls as indicators of the origin of lava fragments emplaced during the 2010 endogenous and exogenous growth in the crater of Volcán de Colima, México. *Bulletin of Volcanology* 79, 47, 1–9. <https://doi.org/10.1007/s00445-017-1130-z>.

associated with the enlargement of the exogenous lava lobe that was first observed in February (see Fig. 16.18E) and continued into March and April.

These results suggest that the analysis of the spectral frequency of seismic signals from rockfalls may provide a way to distinguish the exogenous and endogenous growth of a lava dome. As was noted by Fink et al. (1990), this can help to monitor more closely the long-term extrusion process. Fig. 16.22 illustrates this option, representing the temporal variations in the number of rockfalls, corresponding to LF and HF seismic signals during January–April 2010, for reconstruction of the eruptive process. The figure shows that rockfalls, generated LF seismic signals, those associated with endogenous lava dome growth, continued throughout the entire period. Infrequent rockfalls, emplaced by the shear lava lobe enlargement, began to appear in January. Intensive loss of material from the lobe’s distal end in the first part of February significantly increased the number of HF rockfalls, reaching their peak rate between February 15 and February 20. During

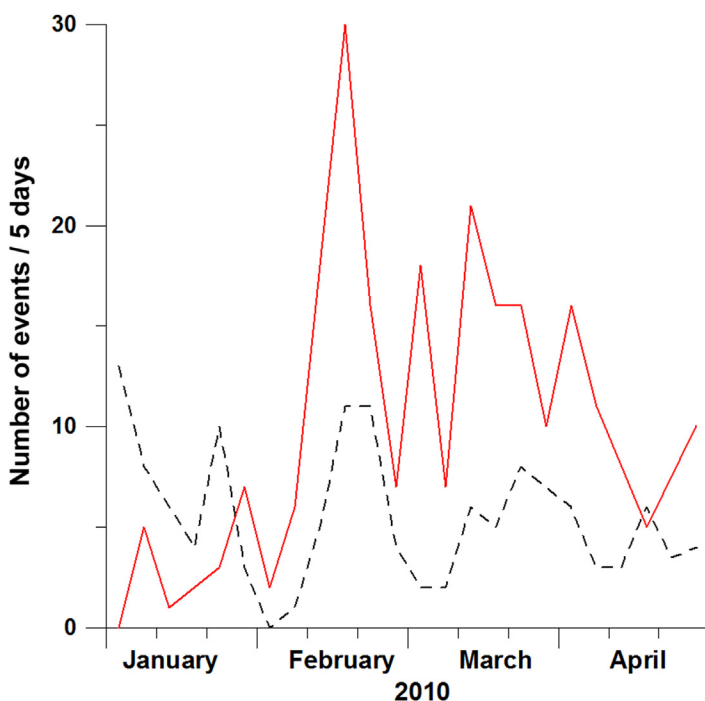


**FIGURE 16.22** Temporal variations in the number of LF (dashed black line) and HF (solid red line) seismic signals of rockfalls recorded at the short-period EZV5 station during January–April 2010. From Zobin, V.M., Tellez, A., Aguilar, J.E., Cruz, K.G. (2017). Seismic signals of rockfalls as indicators of the origin of lava fragments emplaced during the 2010 endogenous and exogenous growth in the crater of Volcán de Colima, México. *Bulletin of Volcanology* 79, 47, 1–9. <https://doi.org/10.1007/s00445-017-1130-z>.



March and April, the number of HF rockfalls decreased, reflecting a gradual lowering in the intensity of the lava lobe enlargement (Zobin et al., 2017).

This chapter demonstrates the great level of informativity of the seismic signals, recorded at different stages of the lava dome eruptions, for detailed analysis of these eruptions' development.

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## RELEVANT WEBSITE

BGVN. Bulletin of Global Volcanic Network. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.

# Swarms of Microearthquakes Associated With Effusive and Explosive Activity at Volcanoes

Swarms of microearthquakes (majority of them with magnitudes of about 0) have been recorded on different volcanoes of the world (e.g., Núñez-Cornú et al., 1994; Orozco Rojas, 1994; Neuberg et al., 1998; Carniel et al., 2006; Iverson et al., 2006; Macedo et al., 2008; Waite et al., 2008; Varley et al., 2010; Zobin et al., 2010). They are associated with effusive and explosive eruptions at volcanoes of all types: basaltic, andesitic, and dacitic. These swarms occur frequently just before large explosions or lava extrusions. While these seismic signals have been given many names (hiccups, pulgas, drumbeats, low-frequency, long-period, and hybrid), in this chapter, the author refers to them as “microearthquakes” and discusses their waveforms, their nature, and their relationship to eruptive events.

## 17.1 WAVEFORMS AND SPECTRA

Fig. 17.1 shows two typical seismograms of the microearthquake swarms recorded at Volcán de Colima. One seismogram (top panel) was recorded just before the beginning of the September 30, 2004 lava extrusion, and another seismogram (bottom panel) was recorded just before the April 30, 2005 large explosion. These seismic signals are numerous, with low amplitude, and form families of similar records (Fig. 17.2).

Three types of microearthquake waveforms were discriminated at Volcán de Colima. The seismic signals and spectral characteristics of Type 1 are shown in Fig. 17.3, together with the waveforms of microearthquakes recorded at volcanoes Stromboli, Ubinas, and Mount St. Helens. The waveforms consist of two phases: an initial low-frequency and small amplitude pulse (between  $t_1$  and  $t_2$ ) and a second, characterized by larger amplitude and higher frequency. The spectral peaks are between 2 and 7 Hz. They are usually referred to as “low-frequency” or “long-period” earthquakes.

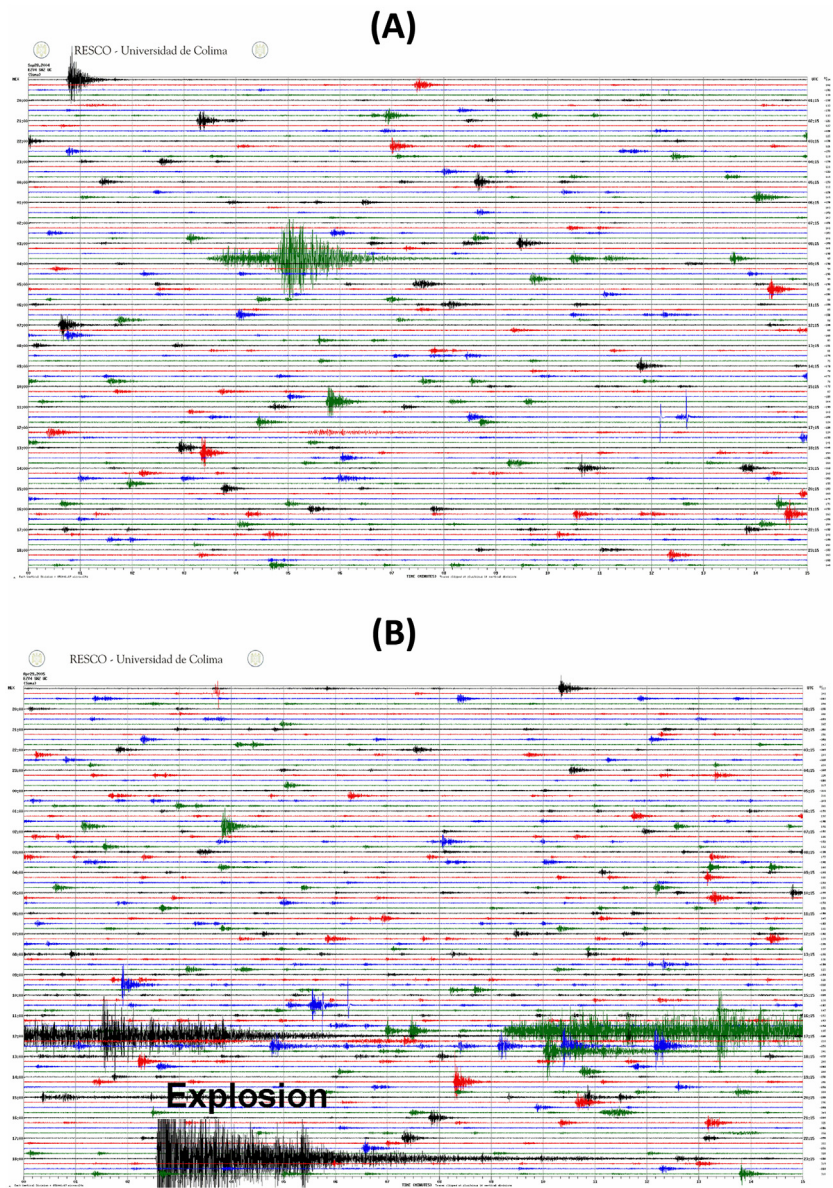


**FIGURE 17.1** Twenty-four hours of seismic records of microearthquake swarms that were observed at Volcán de Colima, Mexico before the September 30, 2004 lava extrusion (A) and before the April 30, 2005 explosion (B). Short-period seismic station EZV4 (velocity, vertical channel), 1.7 km from the crater. *Courtesy Colima Volcano Observatory.*

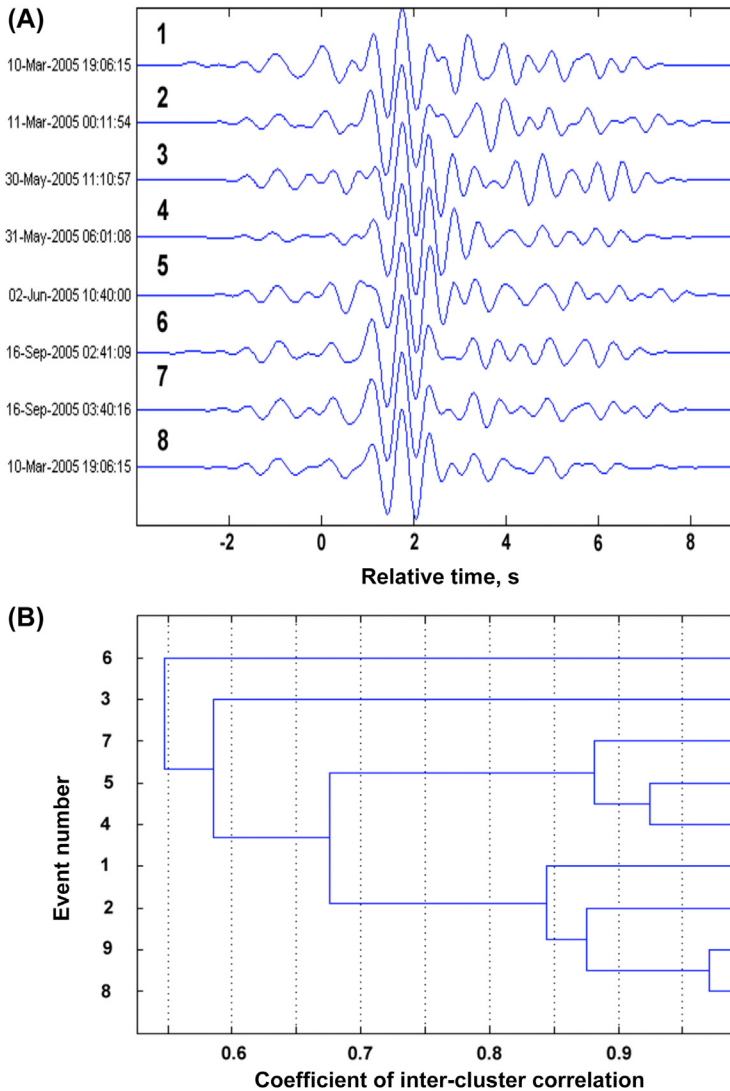


**FIGURE 17.2** Eight seismic records (velocity, vertical component) of the microearthquakes family at Volcán de Colima with adjusted relative time (A) and the intercorrelation between them (B) are shown. The coefficients of intercluster correlation for these eight records, shown in B and marked with the same numbers as in A, vary between 0.55 and 0.95. The majority of the seismic records (5–8) are characterized by the coefficients of correlation between 0.85 and 0.95, which indicates a high level of correlation between these seismic records. From Zobin, V.M., Melnik, O.E., Gonzalez, M., Macedo, O., Bretón M., 2010. Swarms of micro-earthquakes associated with the 2005 Vulcanian explosion sequence at Volcán de Colima, México. *Geophysical Journal International* 182, 808–828.

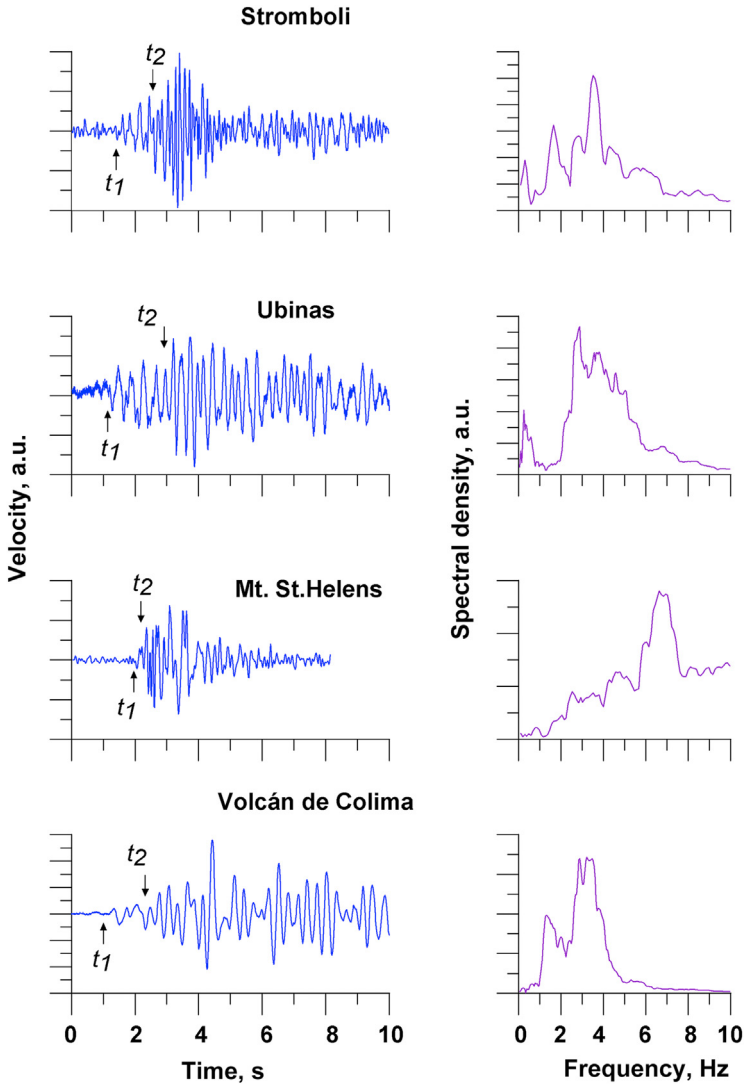


**FIGURE 17.3** Seismograms (velocity, vertical component) of microearthquakes of Type 1 (long period) recorded at volcanoes Stromboli (February 24, 2006, 300 m from the crater, broadband record), Ubinas (May 29, 2006, 2.5 km from the crater, broadband record), Mount St. Helens (March 8, 2005, a few hundred m from the crater, short-period record), and Volcán de Colima (May 14, 2005, 1.7 km from the crater, short-period record) are shown on the left side of the figure and corresponding Fourier spectra of the full records are shown on the right side. Arrows show the first pulse of the seismic record. *Courtesy Sezione di Catania, Istituto Nazionale di Geofisica e Vulcanologia, Instituto Geofísico de Perú, Arequipa, and Colima Volcano Observatory, respectively.*

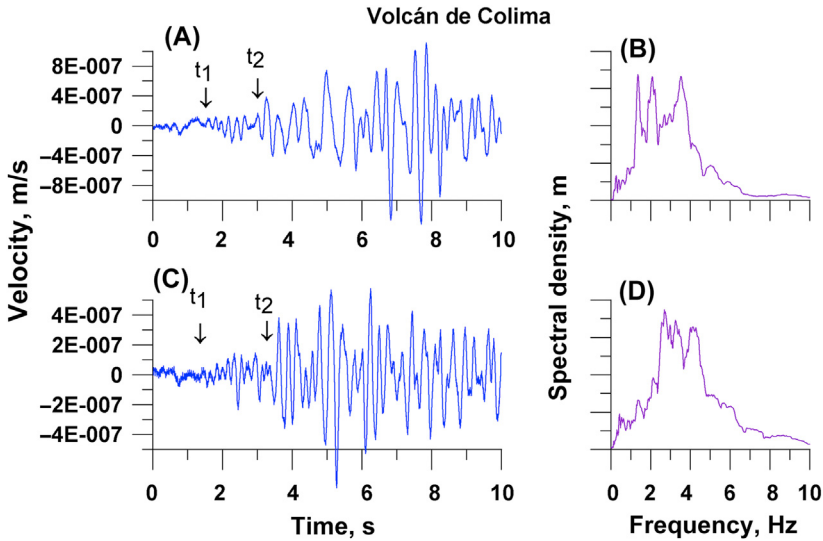


**FIGURE 17.4** Seismograms (velocity, vertical component) of microearthquakes of Type 2 (hybrid) recorded at Volcán de Colima (A, C) during the 2005 activity and corresponding Fourier spectra of the full records (B, D). Arrows show the first pulse of the seismic record. *Courtesy Colima Volcano Observatory.*

The waveforms of microearthquake of Type 2 are shown in Fig. 17.4. The waveforms consist of two phases: an initial high-frequency and small amplitude pulse (between  $t_1$  and  $t_2$ ) and a second, characterized by larger amplitude and lower frequency. The spectral peaks are between 3 and 4 Hz. These waveforms are usually referred as “hybrid” earthquakes.

The waveforms of the microearthquake of Type 3 recorded at Volcán de Colima are shown in Fig. 17.5. These waveforms are characterized by a spindle-shaped seismic signal with a gradual increase and decrease in amplitude, often without a clear signal onset. They may be long (Fig. 17.5A) or short (Fig. 17.5C). The spectral peaks are between 2 and 4 Hz.

## 17.2 STRUCTURE OF MICROEARTHQUAKE SWARMS

Martini et al. (2007) studied the microearthquake swarms recorded before and after explosions that were observed during the 2007 eruption at Stromboli volcano. They showed that the swarms consisted of two types of signals defined as long-period events and hybrid events (Fig. 17.6).

The swarms of microearthquakes (named “drumbeats,” see Chapter 16), that were observed during the 2004–05 effusive–explosive activity at Mount St. Helens volcano, consisted of repetitive long-period (peak frequencies in the 2–3-Hz range) and hybrid (peak frequencies in the 8–16 Hz



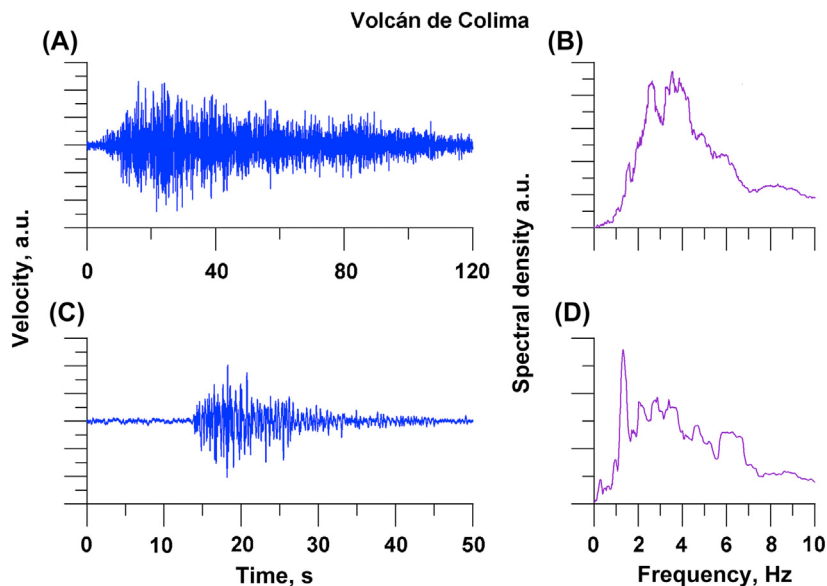


**FIGURE 17.5** Seismograms (velocity, vertical component) of microearthquakes of Type 3 (spindle shaped) recorded at Volcán de Colima (A, C) during 2005 activity and corresponding Fourier spectra of the full records (B, D). *Courtesy Colima Volcano Observatory.*

range) seismic signals which preceded and accompanied the extrusions of several dacite spines. The first swarm of these microearthquakes that occurred on October 16, 2004, is shown in Fig. 17.7. Most of these seismic signals occurred in families of similar events (Fig. 17.8).

Moran et al. (2008a) noted also the occasional occurrence of larger, or “big” ( $M > 2$ ,  $M_{\max} = 3.4$ ) shallow earthquakes within the microearthquake swarms (Fig. 17.9A). Despite their size, the larger earthquakes were similar in several ways to the drumbeat events. Like the drumbeats, they were classified as either long-period or hybrid events. Spectra for drumbeats and the larger events were also similar (Fig. 17.10). Epicenters of larger events were located within a small area of  $1 \text{ km} \times 1 \text{ km}$  centered on the vent; they clustered more densely than drumbeat locations (Fig. 17.9B). The part of small earthquakes was represented with rockfall seismic signals (Moran et al., 2008a,b).

Zobin et al. (2010) showed that the microearthquakes of the first and the second types, which were observed during the 2005 explosive activity at Volcán de Colima and described in Section 16.1, represented 84%–89% of a total of 966 identified events while the microearthquakes of the third type represented from 11% to 16%.

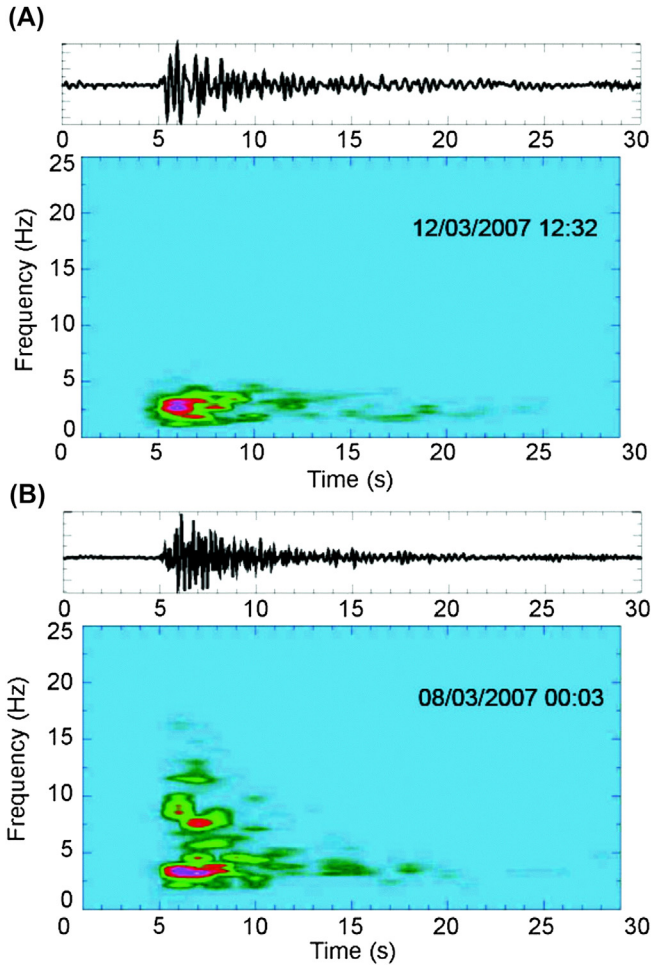


**FIGURE 17.6** Seismic signals (velocity, vertical component) and corresponding spectrograms of long-period (A) and hybrid (B) microearthquakes recorded at Stromboli volcano by broadband seismic station during 2007 activity. *From Martini, M., Giudicepietro, F., D'Auria, L., Esposito, A.M., Caputo, T., Curciotti, R., De Cesare, W., Orazi, M., Scarpato, G., Caputo, A., Peluso, R., Linde, A., Sacks, S., 2007. Seismological monitoring of the February 2007 effusive eruption of the Stromboli volcano. Annals of Geophysics 50, 775–788 with modifications.*

### 17.3 MICROEARTHQUAKE SWARMS IN ERUPTION PROCESS

As was mentioned previously, microearthquake swarms have been observed during effusive and explosive activity at volcanoes of different types all over the world.



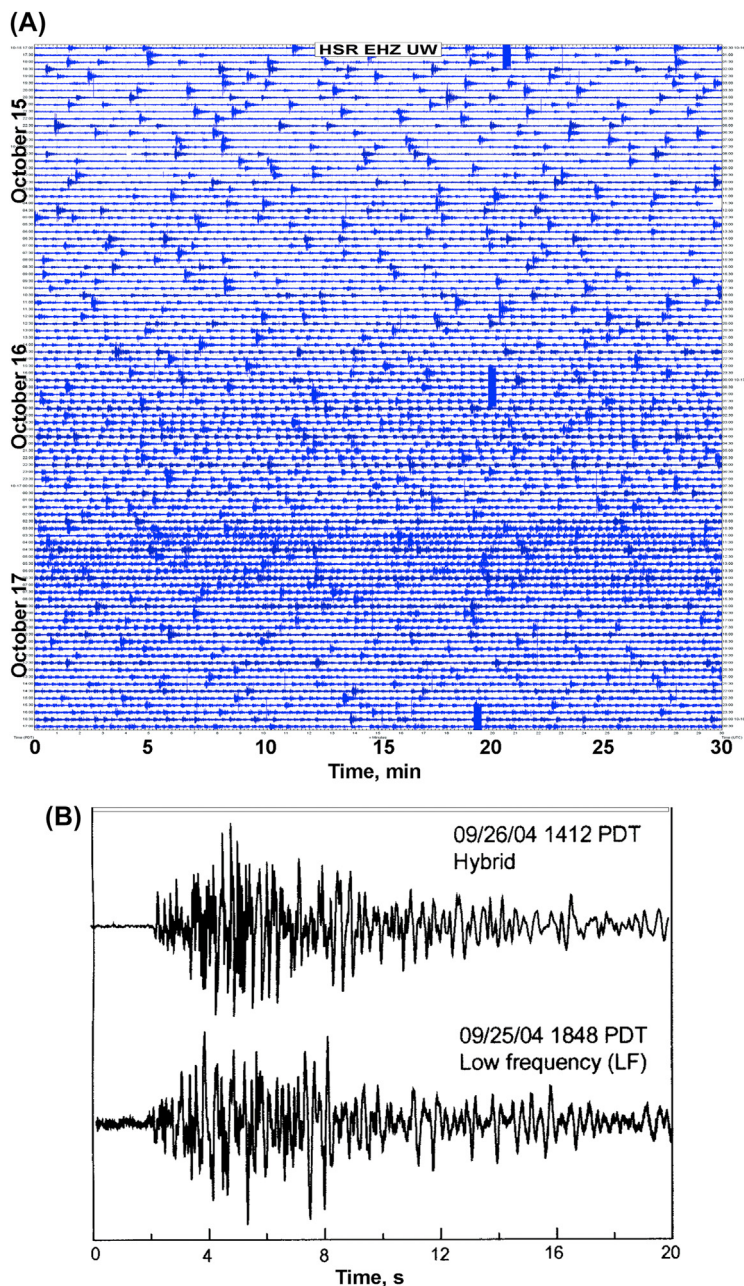


**FIGURE 17.7** Short-period seismic records (velocity, vertical component; 2.5 km from the crater) of the microearthquake swarm that occurred at Mount St. Helens from October 15, 2004, to October 17, 2004 (A) and characteristic seismic records (B) of long-period and hybrid events composing this swarm. After Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*, U.S. Prof. Paper 1750, Washington, pp. 27–60.

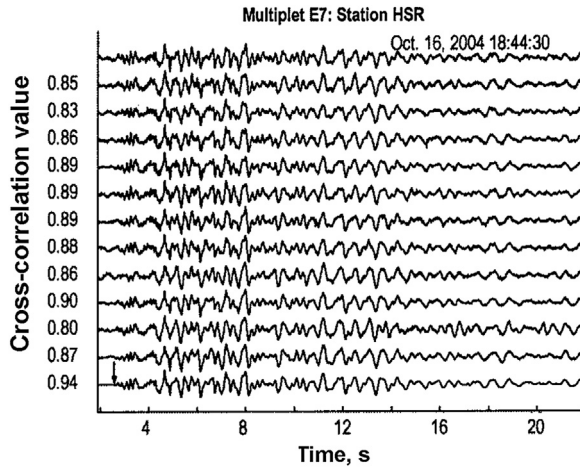


**FIGURE 17.8** Example of a family (multiplet) of waveforms of microearthquakes occurring within the October 16 swarm at Mount St. Helens (see Figure 16.7). The top trace is a seed event for the multiplet, and the last trace is a stack of all waveforms within the multiplet at the given station. The date and time on the first trace are when the seed event occurred. Arrow in the bottom trace shows the first arrival with polarity (up is compressional, and down is dilatational). The cross-correlation value of each trace with respect to the seed event is shown on the y-axis. Waveforms are unfiltered and chosen at constant time intervals within the lifespan of their respective multiplets. After Thelen, W.A., Crosson, R.S., Creager, K.C., 2008. *Absolute and relative locations of earthquakes at Mount St. Helens, Washington, using continuous data; implications for magmatic processes*. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*, U.S. Geological Survey Professional Paper 1750, pp. 71–95.

### 17.3.1 Kizimen Volcano, Kamchatka

The 2376-m-high Kizimen is an isolated, conical andesitic stratovolcano situated in the Eastern Kamchatka, Russia. The latest eruptive cycle began about 3000 years ago with a large explosion and was followed by lava-dome growth lasting intermittently about 1000 years. Only a single explosive eruption, during 1927–28, had been recorded in historical time. New volcanic unrest began in July 2009 with the swarms of volcano-tectonic earthquakes. Beginning in October 2010, strong gas-and-steam activity was recorded. Strong explosions with the ash plumes rising up to 6–7 km above sea level (a.s.l.) were observed during December 2010–May 2011. Numerous pyroclastic flows and a small lava flow were formed (Malik and Ovsyannikov, 2011).

A sequence of microearthquakes began on May 12, 2011, after the large explosion of May 2 (Pavel Firstov, personal communication). Fig. 17.11 shows the sequence of microearthquakes recorded by the short-period seismic station at a distance of 2.6 km from the crater. These short (duration of

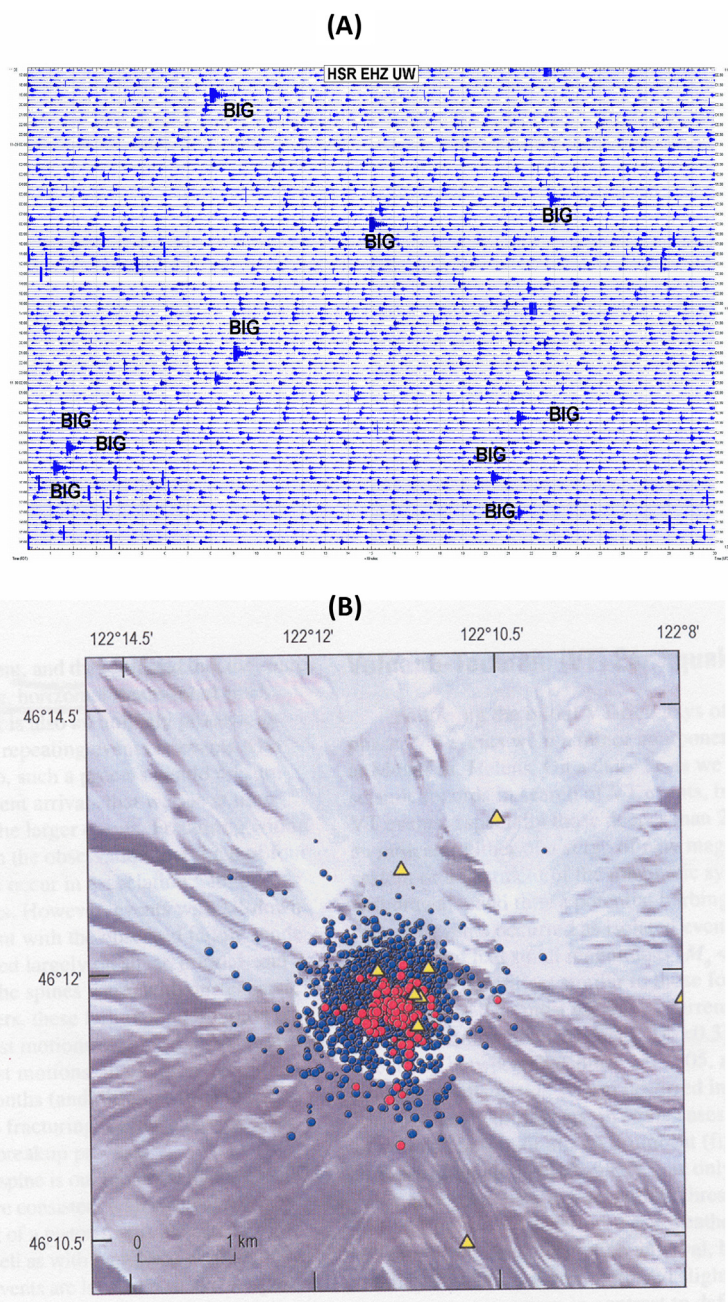


**FIGURE 17.9** Short-period seismic record for 48 h between 17:00 PST on November 28, 2004, and 17:00 PST on November 30, 2004, showing 10 “big” earthquakes occurring along with regular drumbeats (A) and the map showing epicenters of drumbeat events (*blue dots*) and impulsive “big” events (*red dots*) occurring between November 1, 2004, and December 31, 2005 (B). Dot sizes vary with magnitude drumbeat locations include events with *P* arrivals on at least seven stations, the nearest station being within 2 km, and an azimuthal gap  $\leq 135$  degrees (all impulsive “big” event locations meet these same criteria). *Yellow triangles* correspond to seismic station locations as of July 2005. After Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*, U.S. Professional Paper 1750, Washington, pp. 27–60.



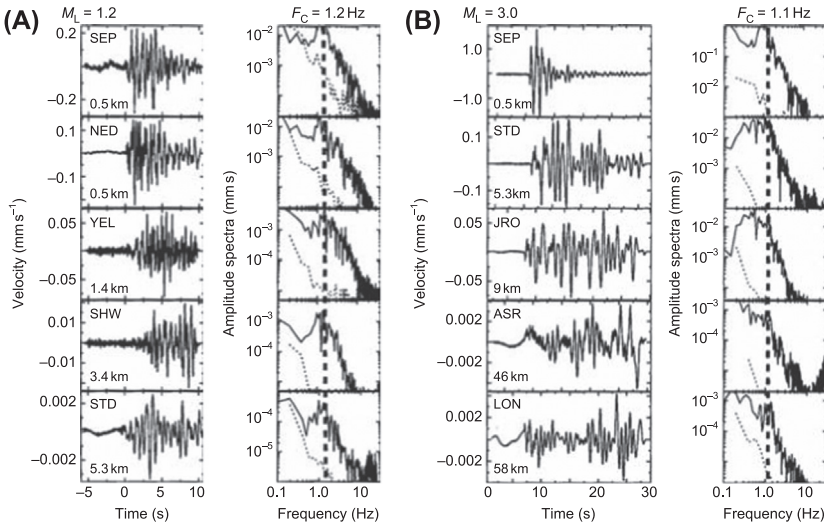


**FIGURE 17.10** Instrument-corrected short-period seismograms and frequency spectra for typical drumbeat (A) and “big” (B) seismic events. Stations are arranged in increasing distance from the vent, with distance indicated for each station. Spectra are for 10.24-s windows starting with *P*-wave arrival. The *thin dotted line* below each spectral plot is a noise estimate based on a 10.24-s window taken before the first arrival. *Vertical thick dashed lines* indicate corner frequencies ( $F_c$ ) for each event. (A) Drumbeat event,  $M_L$  1.2, on August 20, 2005, at 13:45 PDT. (B) “Big” event,  $M_L$  3.0, on March 26, 2005, at 19:28 PST. From Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008a. *Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006*. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*, U.S. Professional Paper 1750, Washington, pp. 27–60.

20–30 seconds) seismic signals are characterized by the long-period initial phase followed by a larger amplitude, high-frequency phase.

### 17.3.2 Stromboli Volcano, Aeolian Islands

The 924-m-high basaltic stratovolcano Stromboli, situated in the Southern Tyrrhenian Sea on the island Stromboli, is an open-conduit volcano characterized by the permanent emission of a gas plume from the craters and persistent activity during at least the last 1000 years (Rosi et al., 2000). The swarms of microearthquakes, represented by hybrid and long-period waveforms, have been observed at different stages of explosive and effusive activity of this volcano.

On April 5, 2003, a powerful explosion covered a large part of the normally tourist-accessible summit area with bombs (Carniel et al., 2006). The event started with a reddish ash emission related to collapses within the craters and was soon replaced by emissions of darker juvenile material from the

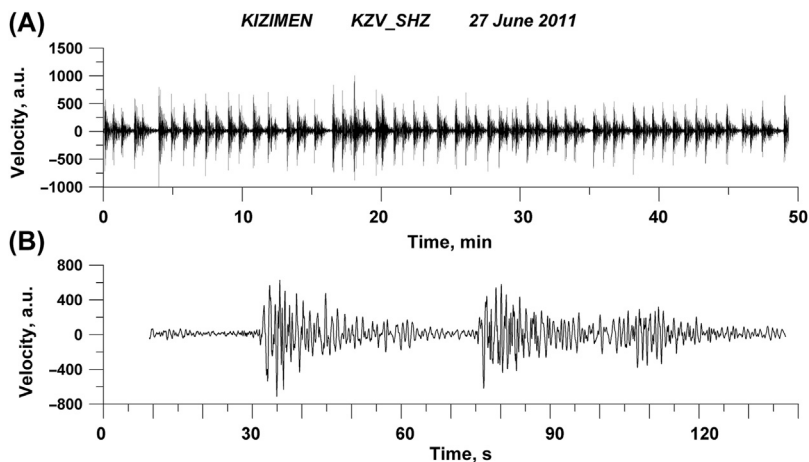


**FIGURE 17.11** Sequence of microearthquakes observed during the 2011 explosive–effusive activity at Kizimen volcano (A) and the detailed seismic record of two of them (B). *Courtesy Institute of Volcanology and Seismology, Kamchatka, Russia.*

northeast crater and then similar emissions from the southwest crater, with the appearance of a dark, mushroom-shaped cloud rising about 1 km above the summit. Bombs, ash, and blocks affected most of the western sector; the volcano top above 700 m a.s.l. was completely covered by bombs, some being meter-sized. This explosion was preceded by a swarm of microearthquakes.

Significant effusive eruption interrupted the persistent explosive activity at Stromboli during the period February 27, 2007–April 2, 2007. Microearthquakes were recorded before, during, and after the effusive episodes (Martini et al., 2007). Fig. 17.12 shows the seismograms filled with microearthquakes recorded just before and after the February 27 effusive eruption. Its onset changed the visible activity of microearthquakes; before the onset of the effusive eruption, they were of larger amplitudes.

Fig. 17.13 shows the location and depth distribution of long-period microearthquakes during the period January 1, 2007–May 22, 2007 (Martini et al., 2007). The temporal range is divided into four intervals: (1) January 1–January 31, (2) January 31–February 26, (3) February 27–March 31, and (4) March 31–May 22. There were no significant changes between (1) and (2) periods (before the effusive phase), whereas a depth increase is evident in (3) and (4) periods (after the beginning of the effusive phase).

### 17.3.3 Mount St. Helens, Cascades

The 2549-m-high dacitic stratovolcano Mount St. Helens began erupting on October 1, 2004, after 18 years of quiescence (see Section 7.1.1). The newly

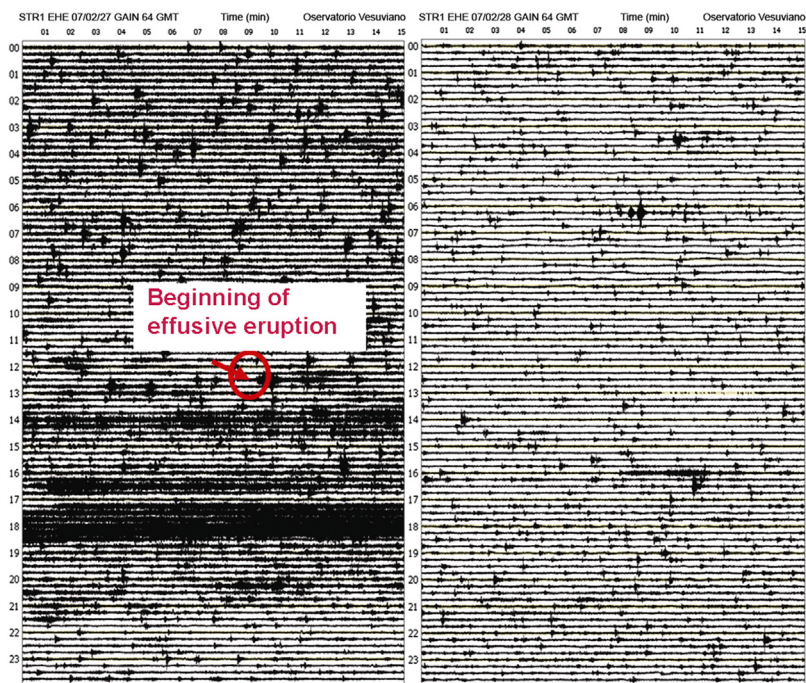


**FIGURE 17.12** Microearthquake swarms observed during February 27, 2007 (left panel) and February 28, 2007 (right panel) at Stromboli volcano just before and after the February 27 effusive eruption. The circle indicates the start of the effusive eruption. From Martini, M., Giudicepietro, F., D'Auria, L., Esposito, A.M., Caputo, T., Curciotti, R., De Cesare, W., Orazi, M., Scarpato, G., Caputo, A., Peluso, R., Linde, A., Sacks, S., 2007. Seismological monitoring of the February 2007 effusive eruption of the Stromboli volcano. *Annals of Geophysics* 50, 775–788.

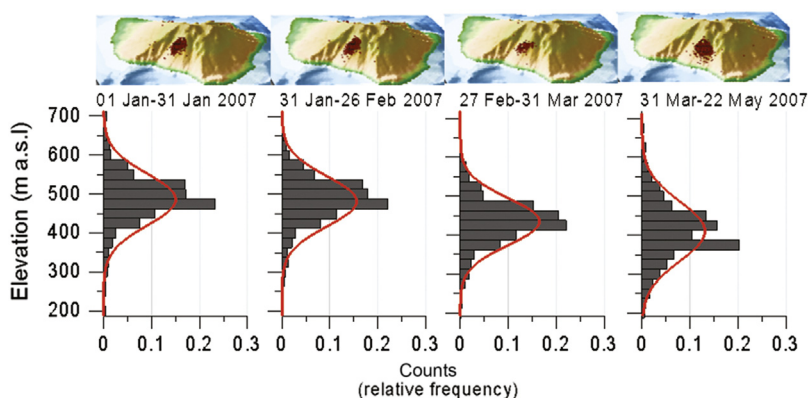


**FIGURE 17.13** The location (above) and the histograms of the hypocentral elevation of microearthquakes (below) before and after the beginning of the effusive eruption (February 27, 2007) at Stromboli volcano. From Martini, M., Giudicepietro, F., D'Auria, L., Esposito, A.M., Caputo, T., Curciotti, R., De Cesare, W., Orazi, M., Scarpato, G., Caputo, A., Peluso, R., Linde, A., Sacks, S., 2007. Seismological monitoring of the February 2007 effusive eruption of the Stromboli volcano. *Annals of Geophysics* 50, 775–788.

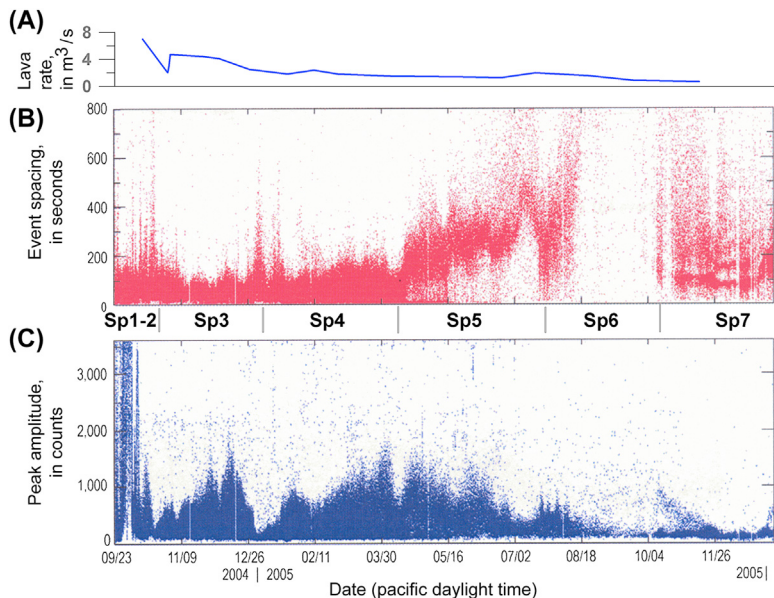


**FIGURE 17.14** Plot showing the variations in the volumetric lava-dome growth rate (A) together with the interevent spacing (B), and peak amplitudes (C) for detected microearthquakes during the 2004–05 eruption at Mount St. Helens volcano recorded at a distance about 3 km from the crater between September 23, 2004, and December 31, 2005. The intervals of the spines 1–7 (Sp1–Sp7) growths are shown. Data from Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Professional Paper 1750, Washington, pp. 27–60 and Vallance, J.W., Schneider, D.J., Schilling, S.P., 2008. Growth of the 2004–2006 lava-dome complex at Mount St. Helens, Washington. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The First Year of Renewed Eruptions at Mount St. Helens, 2004–2006*, U.S. Geological Survey Professional Paper, 1750, pp. 169–208.

extruded dacite formed a series of spines beginning on October 11. These spines were repeatedly formed and partially destroyed (Iverson et al., 2006). A sequence of explosions was recorded during 2004–05 (Moran et al., 2008a,b).

A series of long-period and hybrid microearthquakes with similar waveforms began to appear from September 26 and accompanied the October 1–October 5 vent-clearing phases of small explosions and the initial stage of the lava-dome growth that commenced on October 11. Beginning from October 16, the microearthquake events appeared to be regularly spaced (1–3 minute intervals) which inspired them to be named “drumbeats.” Significant variations in drumbeats’ size and spacing (Fig. 17.14B and C) were observed during the 2004–05 growth of lava spines 1–7, indicating the correlation of the drumbeats characteristics with changes in the style of

extrusion at the surface. At the same time, changes in drumbeat character did not correspond to variations in magma flux at the conduit (Fig. 17.14A), indicating that drumbeat size and spacing may be more a function of the mechanics of extrusion than of the extrusion rate (Moran et al., 2008a).

During the 2004–05 eruption at Mount St. Helens, six explosions were recorded: four of them occurred during the vent-clearing phase from October 1, 2004, to October 5, 2004, before the lava-dome growth, and two explosions occurred on January 16, 2005 and March 8, 2005, during the dome-building phase (Moran et al., 2008b). The ends of two “big” earthquake clusters coincided with explosions on January 18, 2005, and March 8, 2005. Fig. 17.15

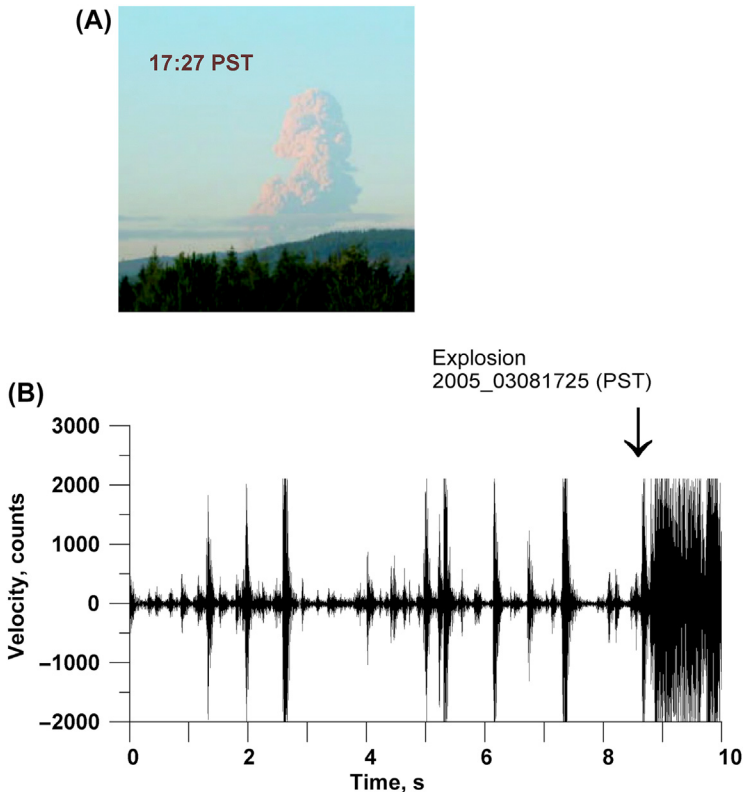


**FIGURE 17.15** The March 8, 2005 explosion at Mount St. Helens volcano (A) and a sequence of drumbeat microearthquakes recorded by a short-period seismic station at a distance of a few hundred meters from the crater just before the explosion (B). Photo is taken from Moran, S.C., McChesney, P.J., Lockhart, A.B., 2008b. *Seismicity and infrasound associated with explosions at Mount St. Helens, 2004–2005*. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), *A Volcano Rekindled: The First Year of Renewed Eruptions at Mount St. Helens, 2004–2006*, pp. 111–127 U.S. Geolog. Surv. Prof. Pap., 1750; the seismic records were provided by Stephan Malone (Pacific Northwest Seismic Network).



shows a sequence of micro- and big earthquakes, preceding the largest of them, the March 8, 2005 explosion. Some clusters of “big” earthquakes from the drumbeat swarms occurred in association with the breakup of lava spines and during the steady extrusion of one of them (Moran et al., 2008a).

### 17.3.4 Ubinas Volcano, Perú

The 5672-m-high andesitic-to-rhyolitic stratovolcano Ubinas has been the most active Peruvian volcano for the last 450 years. Its historical activity, documented since the 16th century, has consisted of intermittent minor-to-moderate explosive eruptions. The recent unrest at Ubinas began on March 25, 2006, with small ash explosions. A new lava dome in the crater began to grow in April 2006. A sequence of large explosions accompanied by the ash columns and reaching 7–8 km a.s.l. occurred from May 2006 to October 2006; the ash eruptions and steam emissions of different intensities continued until at least January 2009 (Macedo et al., 2008). Of the 134 explosions, 43 were preceded by sequences of microearthquakes (Macedo et al., 2008). Fig. 17.16 shows the fragment of a sequence of microearthquakes preceding the May 29, 2006 explosion. The microearthquakes form families of similar waveforms, with dominant frequencies between 2.8 and 3.6 Hz (Fig. 17.17).

### 17.3.5 Volcán de Colima, Mexico

Microearthquake sequences at andesitic Volcán de Colima were recorded firstly during its 1991 eruption (Fig. 17.18). This earthquake sequence continued for 1 day only, 28 days after the beginning of lava extrusion. The dominant frequency of their seismic records was about 3 Hz (Núñez-Cornú et al., 1994; Orozco Rojas, 1994). No microearthquake sequences were

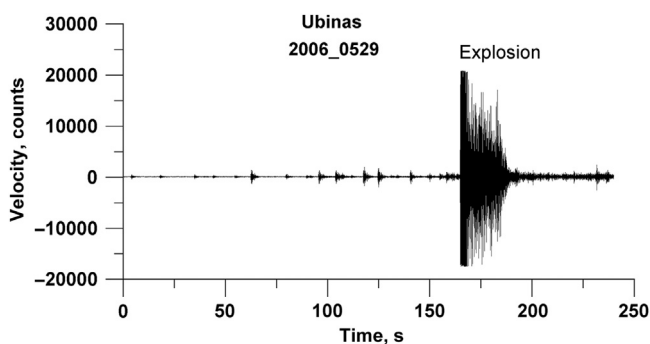


**FIGURE 17.16** The short-period record (velocity, vertical component) of the sequence of long-period microearthquakes before the May 29, 2006 explosion at Ubinas volcano, Peru (2.5 km from the crater). The seismic records were provided by Orlando Macedo (Instituto Geofísico del Perú). Data from Instituto Geofísico del Perú.

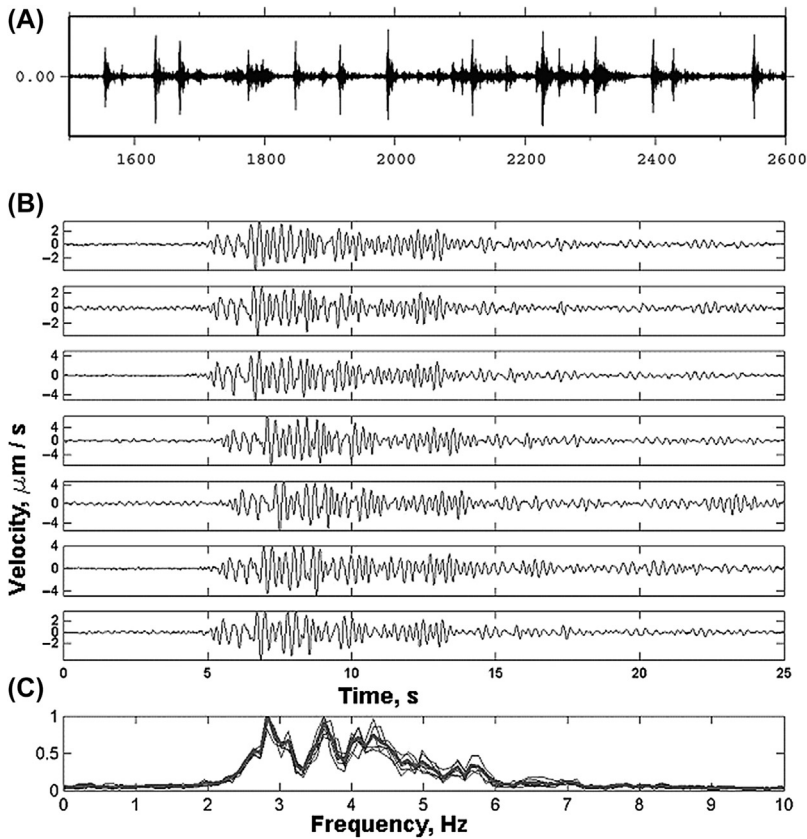


**FIGURE 17.17** A family of the 2006 long-period microearthquakes of Ubinas volcano recorded by short-period seismic station (velocity, vertical component) at a distance of 2.5 km from the crater. (A) A sequence of microearthquakes; (B) presentation of the recorded waveforms as individual events; (C) Fourier spectra of the events presented in (B). *From Macedo O., Métaxian J.-Ph, Taïpe E., Ramos D., 2008. Actividad sismo-volcánica asociada a la erupción del volcán Ubinas en 2006–2008. In: XIII Congreso Latinoamericano de Geología, Lima, Perú, 29 Setiembre – 3 Octubre, 2008.*

recorded before or during the 1994 phreatic explosion (Jiménez et al., 1995). During the 1999–2005 sequence of large Vulcanian explosions at Volcán de Colima, the appearance of microearthquake swarms became regular. The microearthquake sequences were recorded a few hours before the largest explosions of 1999 on February 10, May 10, and July 17 (Zobin et al., 2002; Vargas-Bracamontes et al., 2009).

Most numerous and long-acting microearthquake activity at Volcán de Colima was observed during the March 2005–September 2005 explosion sequence, when 15 explosions with energy  $\geq 10^{11}$  J were recorded (Zobin et al., 2006); all these explosions were preceded and accompanied by

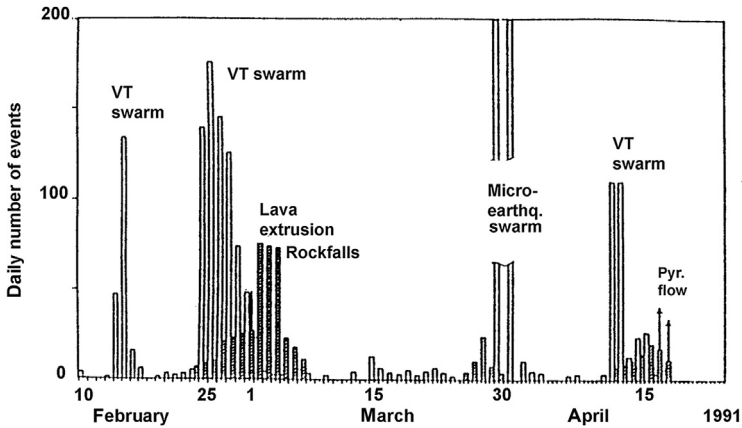


**FIGURE 17.18** Seismic activity at Volcán de Colima during February 1994–April 1994. One-day-duration swarm of microearthquakes was recorded 28 days after the beginning of the lava extrusion. *From Orozco Rojas, J., 1994. Analysis of the peculiar swarm of low-frequency earthquakes detected at Colima volcano, Mexico. Individual Studies by Participants in the International Institute of Seismology and Earthquake Engineering, Tokyo, vol. 30, 95–109.*

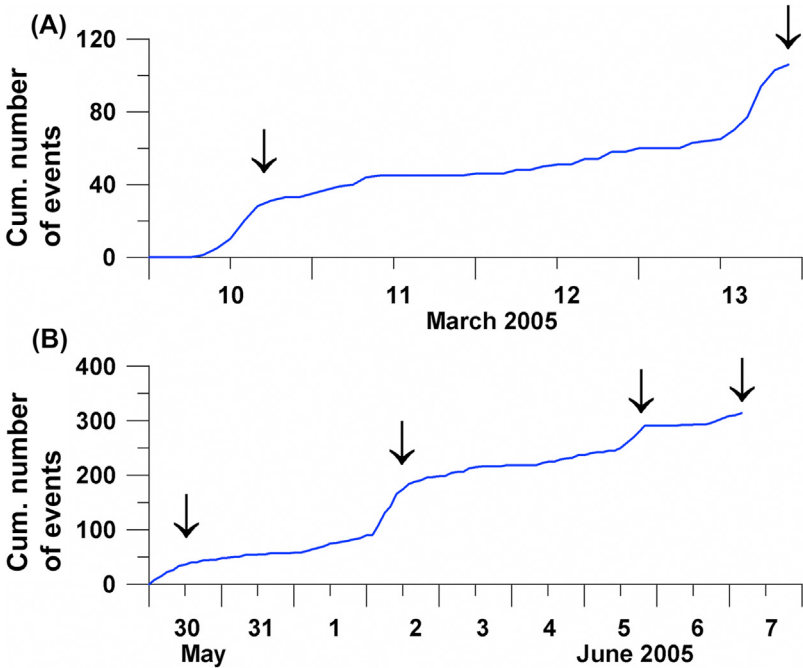


**FIGURE 17.19** Temporal variations in the cumulative number of microearthquakes during the March 2005 (A) and May 2005–June 2005 (B) explosive sequences at Volcán de Colima. The moments of explosions are shown by the arrows. *Courtesy Colima Volcano Observatory.*

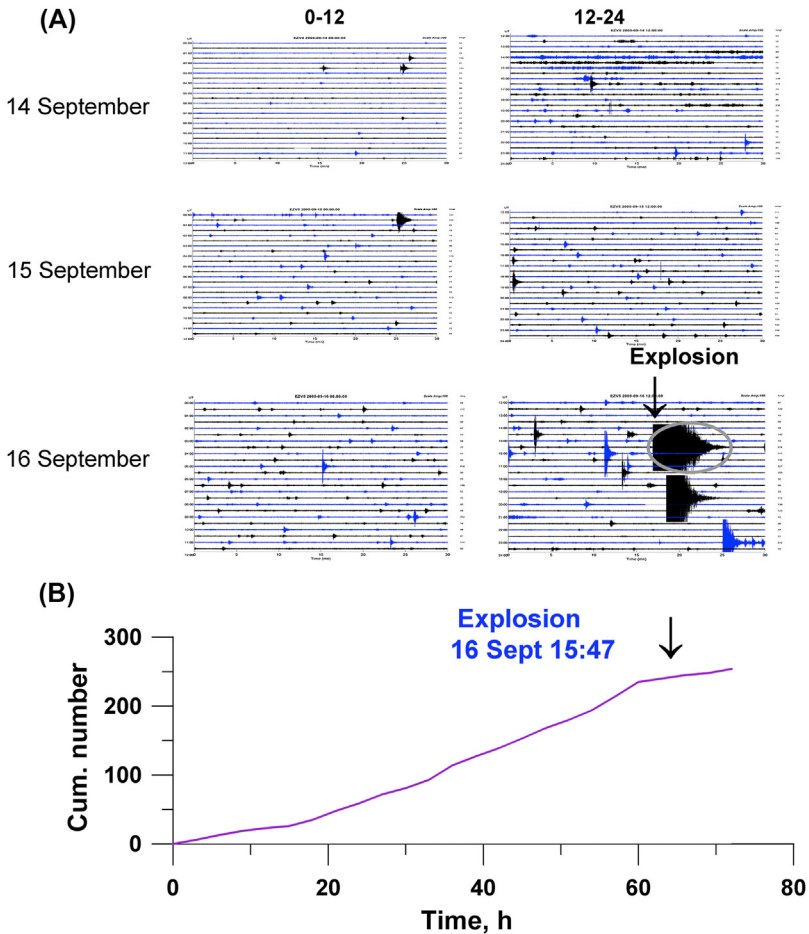


**FIGURE 17.20** The 12-h seismic records (short-period seismic station EZV5, 4 km from the crater) during 3 days of microearthquake activity at Volcán de Colima before the September 16, 2005 explosion (A) and the curve of the cumulative number of microearthquakes (B). The left panel in (A) presents the seismograms recorded during the first 0–12 h and the right panel, during 12–24 h. The moment of the explosion is shown by an arrow. The seismic record of the explosion is marked by an ellipse. *Courtesy Colima Volcano Observatory.*

microearthquake swarms over several days. Fig. 17.19 shows the cumulative number of microearthquakes two main explosive sequences during March 2005 and May 2005–June 2005. It can be seen that every time, some hours before the large explosions, the cumulative number of events sharply increased. The duration of these preliminary intervals varied from 6 to 14 hours without any correlation with the energy of the forthcoming explosions. Between the large explosions, the rate of microearthquake occurrence decreases.

Fig. 17.20A shows the details of the development of a swarm of microearthquakes before an explosion. It shows a set of seismograms recorded at station EZV5 from September 14, 2005, to September 16, 2005, before the explosion of September 16. The moment, when the microearthquake sequence began on September 14, is seen clear enough in the seismogram of the 12–24 hours period. In total, 243 events were recorded. Their rate of appearance increased about 17 hours after the beginning of the sequence (Fig. 17.20B). The cumulative number continued to grow gradually until the moment of the explosion that occurred on September 16 at 15:47, about 60 hours after the beginning of the microearthquake sequence. The rate of the number of microearthquakes strongly decreased after the explosion. The sequence was composed of different types of small seismic signals; the majority of microearthquakes had their durations between 20 and 30 seconds. The microexplosive events were more numerous (154 events) and represented 88% of the recognized events.

The development of the microearthquake swarm that occurred before the September 30, 2004 lava extrusion at Volcán de Colima is shown in Fig. 17.21. The swarm was recorded from September 26, 2004, to September 30, 2004 (Fig. 17.21A). The rate of occurrence of microearthquakes slightly increased about 18 hours after the beginning of the sequence (Fig. 17.21B). The cumulative number continued to grow gradually until the beginning of the lava extrusion about 70 hours after the beginning of the microearthquake sequence.

## 17.4 NATURE OF MICROEARTHQUAKES

The repetitive waveforms of microearthquakes and their occurrence in families imply a nondestructive, repetitive source mechanism that has a stationary source location. The type of these mechanisms is discussed.

### 17.4.1 Similarity Between the Microearthquake Waveforms and the Seismic Signals Well Associated With the Volcanic Events

Simultaneous video images and seismic records of eruptive processes have allowed the characterization of seismic signals described previously as microearthquakes (Zobin et al., 2010). Through comparison between the seismic signals, generated by volcanic explosions and rockfalls, and the simultaneous video images (Fig. 17.22A and B, Fig. 17.23A and B, and Fig. 17.24A and B), it was possible to characterize the events. The comparison of identified seismic records of explosions with the seismic records of microearthquakes of the first and second types shows that both waveforms have either a low-frequency first pulse (Fig. 17.22; long-period events) and/or a high-frequency first pulse (Fig. 17.23; hybrid events), which may be associated with explosion-type sources. Martini et al. (2007) also noted the similarity among the waveforms and spectrograms of volcanic explosions

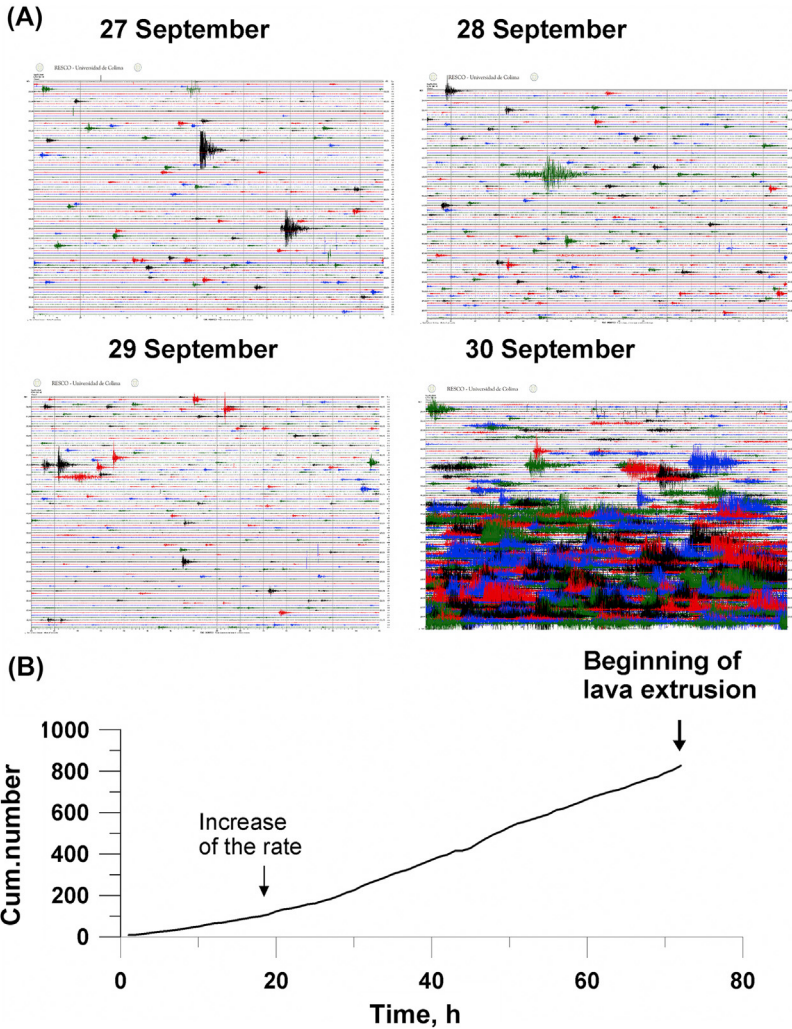


**FIGURE 17.21** The 24-h seismic records (short-period seismic station EZV4, 1.7 km from the crater) during 4 days of microearthquake activity at Volcán de Colima (A) and the curve of the cumulative number of microearthquakes (B) before the September 30, 2004 lava extrusion. The beginning of extrusion is marked by a swarm of rockfall earthquakes on the seismogram of September 30 and is shown by an arrow on the curve. *Courtesy Colima Volcano Observatory.*

(Fig. 17.25A) and long-period microearthquakes (Fig. 17.25B) occurring at Stromboli volcano. They suggested that these two types of seismic signals are actually related to the same source mechanism (gas-slug bursting).

The difference in the spectral content of the initial pulses may be related to different types of triggers of microexplosions within the crack system,



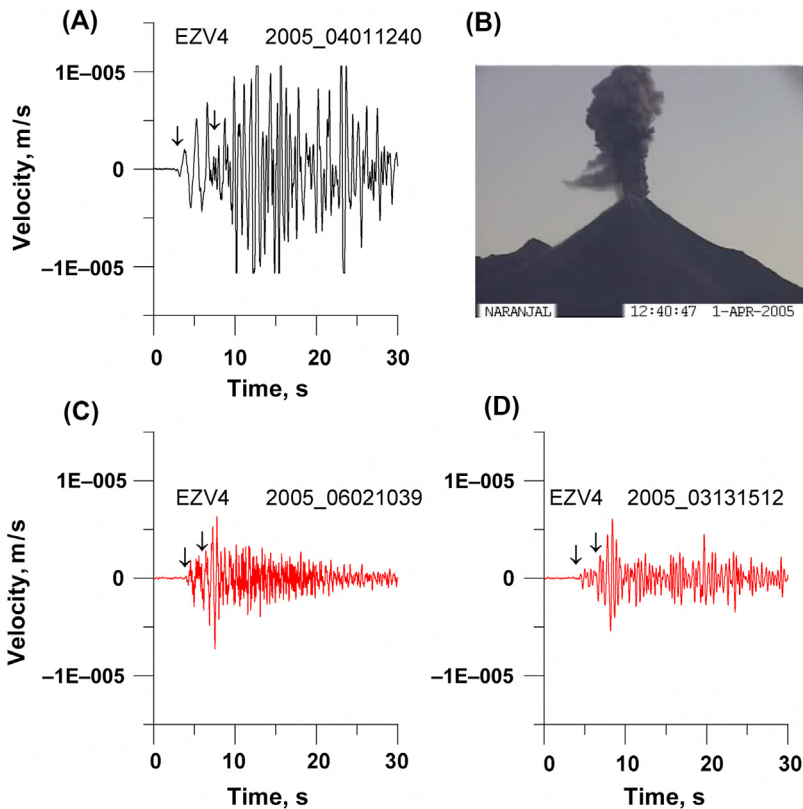


**FIGURE 17.22** The comparison of the seismic records of an explosion (A), identified with a video image (B), with two records of microearthquakes of the first type (C, D) occurring at Volcán de Colima. The seismic station was situated at a distance of 1.7 km from the crater. Code is yyyy\_mmddhhmm. Arrows show the duration of the first pulse. The velocity records are unfiltered and corrected for instrument response. *Courtesy Colima Volcano Observatory.*

which generates microearthquakes of both types. It may be, for example, the vibration of a crack, containing magmatic fluids, during its filling by gas and steam exsolved from ascending magma. This process is comparable with low-frequency vibrations generated by gas bubbles flowing upward within the pipe (foam model; see Section 3.2.1). In another turn, the initial high-frequency pulse may be the result of brittle fracturing of high-temperature rocks (Tuffen et al., 2003, 2008; see Section 3.2.1) surrounding the fluid-filled crack. The following explosion (or gas-slug burst) would be the result of a piston-like action of the gas-steam jet escaping with a high velocity from the crack. This volumetric process produces the second high-amplitude, high-frequency pulse whose spectral amplitude gives the energy of an explosive event.

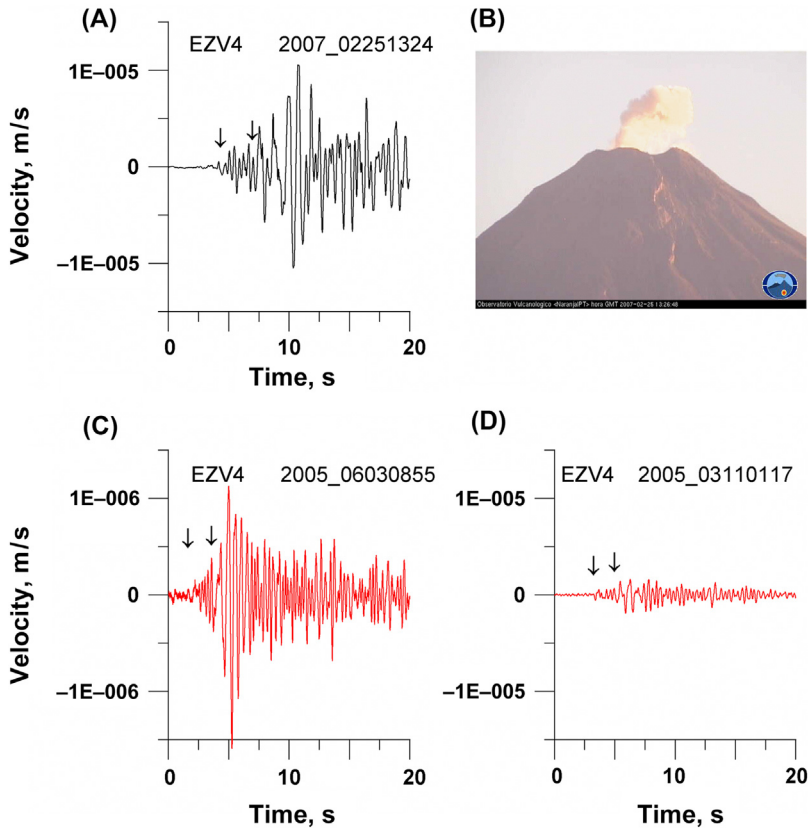


**FIGURE 17.23** The comparison of the seismic records of an explosion (A), identified with a video image (B), with two records of microearthquakes of the second type (C, D) occurring at Volcán de Colima. The seismic station was situated at a distance of 1.7 km from the crater. Code is yyyy\_mmddhhmm. Arrows show the duration of the first pulse. The velocity records are unfiltered and corrected for instrument response. *Courtesy Colima Volcano Observatory.*

A comparison between the seismic records of a rockfall with the seismic records of the microearthquakes of the third type (Fig. 17.24) shows a similar spindle-type waveform. Therefore, the third-type microearthquakes may be considered as produced by small rockfalls.

#### 17.4.2 Quantification of Microearthquakes

According to these conclusions, the methodologies developed for quantification of the seismic signals associated with explosive events (Chapter 14) and rockfalls (Chapter 13) were applied to quantify the 2005 microearthquakes occurring at Volcán de Colima (Zobin et al., 2010).



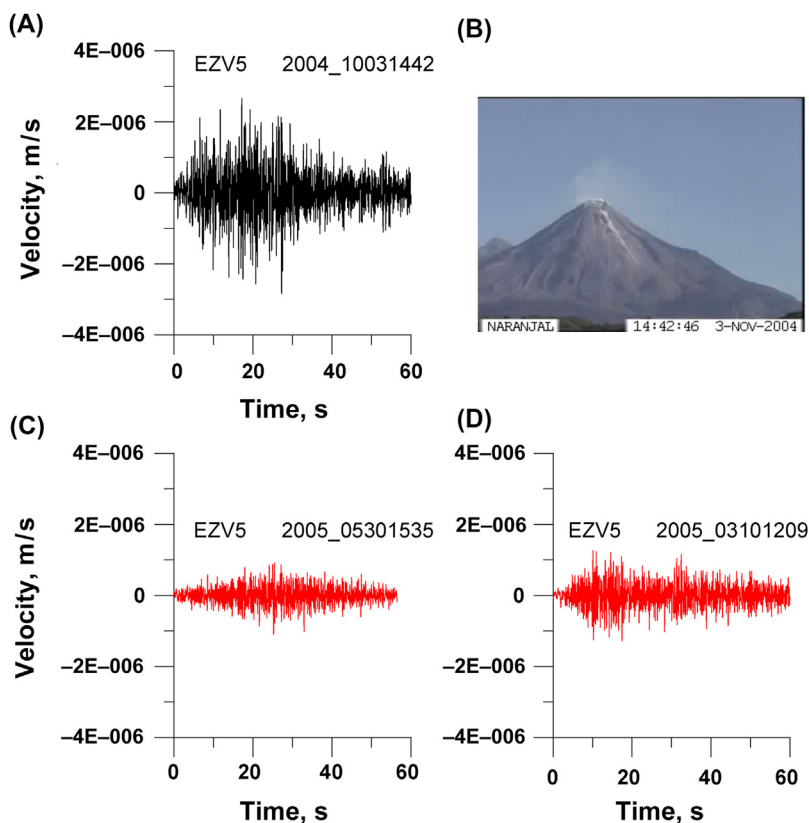


**FIGURE 17.24** A comparison between the seismic records of a rockfall (A), identified with a video image (B), and two records of microearthquakes of the third type (C, D) occurring at Volcán de Colima. The seismic station was situated at a distance of 1.7 km from the crater. Code is yyyy\_mmddhhmm. The velocity records are unfiltered and corrected for instrument response. *Courtesy Colima Volcano Observatory.*

### 17.4.2.1 Quantification of Microexplosions

The energy of 418 microexplosions, calculated from Fourier spectra of high-amplitude, high-frequency pulse, ranged over about four orders varying from  $1 \times 10^4$  to  $7.2 \times 10^7$  J (Zobin et al., 2010). Fig. 17.26 shows the distributions of the energy for the groups of microexplosions, occurring from March 2005 (curve 1) and May 2005 to June 2005 (curve 2). The Kolmogorov–Smirnov test was applied (Fig. 17.26B) to show that these two distributions belong to the same sample with the characteristic mean values and, consequently, occur in the same conditions of their generation. The maximum difference between the two cumulative distributions,  $D_{n1,n2}$ , was 0.1644. We checked the significance of  $D$  at the level of  $\alpha = 0.01$  ( $k_1 - \alpha = 1.628$ ) obtaining the critical value of 0.1830, which is greater than the observed  $D_{n1,n2}$ . It allows

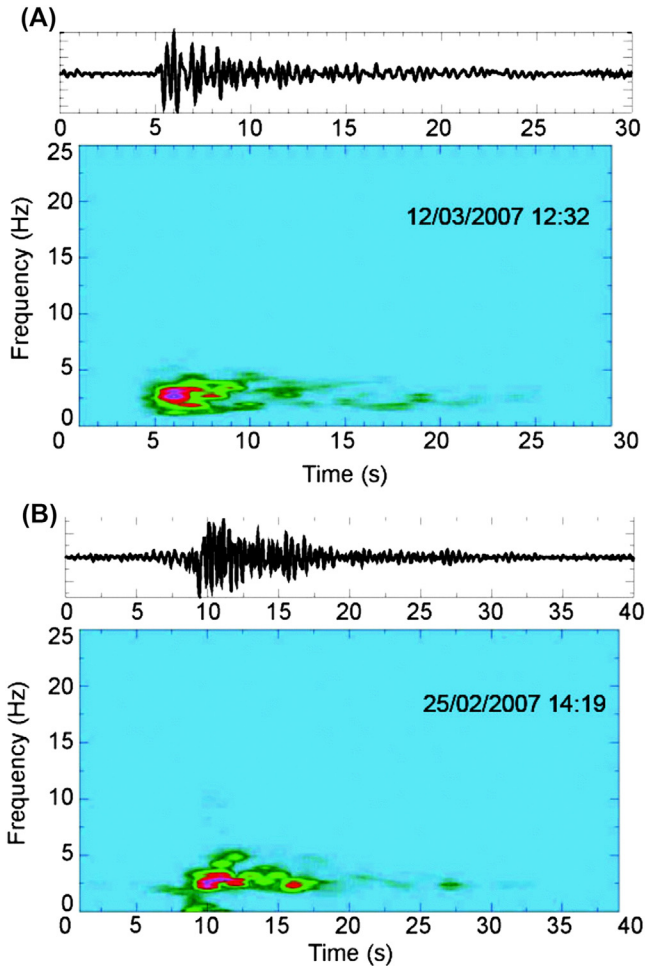


**FIGURE 17.25** Seismic signals (velocity, vertical component) and corresponding spectrograms of explosion earthquake (A) and long-period microearthquake (B) recorded at Stromboli volcano by broadband seismic station during the 2007 activity. *After Martini, M., Giudicepietro, F., D'Auria, L., Esposito, A.M., Caputo, T., Curciotti, R., De Cesare, W., Orazi, M., Scarpato, G., Caputo, A., Peluso, R., Linde, A., Sacks, S., 2007. Seismological monitoring of the February 2007 effusive eruption of the Stromboli volcano. Annals of Geophysics 50, 775–788.*

us to consider that both databases of microexplosions belong to the same sample and may be considered as having the same nature of generation.

The comparison of energy distributions of microexplosions with the energy distributions obtained for 236 small explosions, associated with the September 2004–November 2004 lava extrusion (Fig. 17.26A, curve 3), demonstrates (Fig. 17.26C) that they, in their turn, significantly differ. The

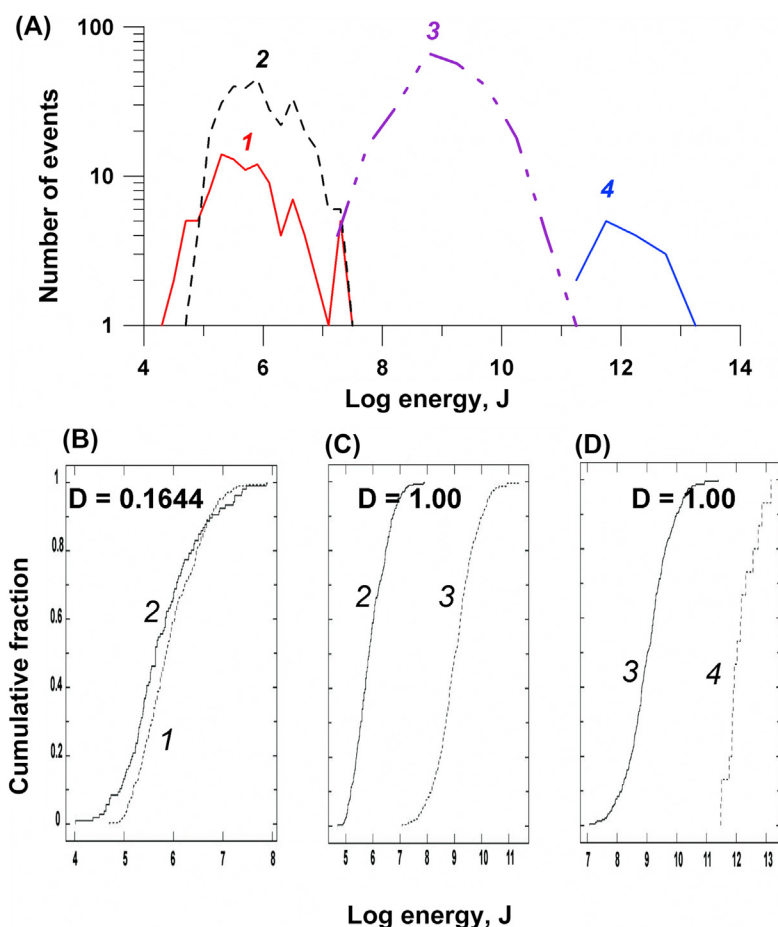


**FIGURE 17.26** Energy distributions (A) obtained for microexplosions occurring at Volcán de Colima during the March 2005 (1) and the May 2005–June 2005 (2) sequences and for small explosions of the November 2004 sequence (3) at Volcán de Colima. (B–D) Kolmogorov–Smirnov test comparison cumulative fraction plots for the sequences 1 and 2, and 2 and 3, respectively. The statistic  $D$  is the maximum vertical deviation between two curves. *Courtesy Colima Volcano Observatory.*

maximum difference between the cumulative distributions,  $D$ , is 1.00 (Fig. 17.26C), which is larger than the 0.01—critical value of 0.1402.

Considering the result (Image 14.27 from Chapter 14) showing the difference in the samples of energy distribution for small and large explosions, this suggests that microexplosions occurred in conditions that differ from those of small and large explosions and may be generated by different processes within the volcanic conduit. The separation between large and small explosions, as was explained in Chapter 14, is at a level corresponding to

$10^{11}$  J and the separation between small and microexplosions is at a level corresponding to  $10^7$  J.

#### 17.4.2.2 Quantification of Rockfall-Type Microearthquakes

Rockfall-type signals generally represent a small part of microearthquake records. These events occur as a result of the partial collapse of active lava domes or flows, or other instabilities on the flanks. The size of the rockfall deposit may be approximately estimated from the duration of the seismic signal (Chapter 13). [Zobin et al. \(2005\)](#) studied the dependence of the duration of short-period seismic records at a distance of 4–7 km from the crater on the volume of the corresponding pyroclastic flow deposit. It was shown that the short-period seismic signals can be divided into three categories based on their duration: short events with durations less than 100 seconds, intermediate events with durations between 100 and 250 seconds, and long events with durations longer than 250 seconds. It was inferred that long events correspond to pyroclastic flows with a mean deposit volume  $\sim 2 \times 10^5 \text{ m}^3$ , and intermediate events represent pyroclastic flows with a mean deposit volume  $\sim 1 \times 10^3 \text{ m}^3$ . Field observations suggest that short events correspond to rockfalls with a mean volume of  $\sim 50 \text{ m}^3$ .

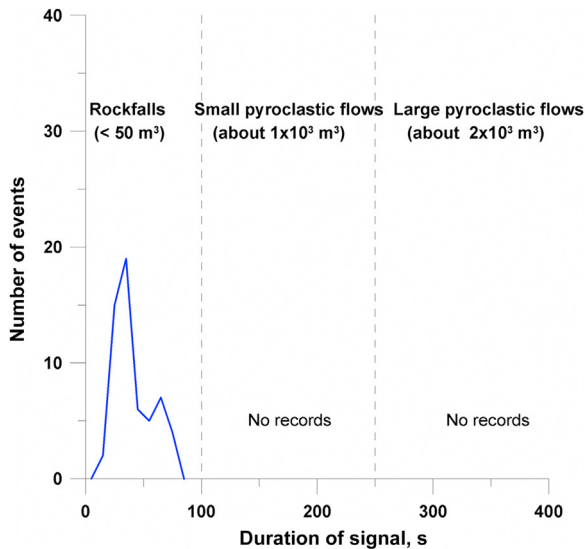


**FIGURE 17.27** Distribution of the durations of rockfall-type microearthquakes occurring at Volcán de Colima during 2005 explosive activity and their position as compared with the durations of short-period seismic signals for different volumes of pyroclastic flows deposits and rockfalls. The mean volume of deposits of pyroclastic flows and rockfalls associated with the seismic records of different durations is shown according to [Zobin et al. \(2005\)](#). From Zobin, V. M., Melnik, O.E., Gonzalez, M., Macedo, O., Bretón M., 2010. Swarms of micro-earthquakes associated with the 2005 Vulcanian explosion sequence at Volcán de Colima, México. *Geophysical Journal International* 182, 808–828.

Fig. 17.27 shows the distribution in durations of the rockfall-type micro-earthquake seismic signals recorded during the 2005 explosive activity at Volcán de Colima (Zobin et al., 2010). All of them were characterized by durations less than 100 seconds, settling them within the category of short events which correspond to rockfalls with a mean volume of  $\sim 50 \text{ m}^3$ .

### 17.4.3 Nature of Microearthquakes Resolved From Waveform Inversion

Waite et al. (2008) modeled the source mechanisms of low-frequency microearthquakes, occurring at Mount St. Helens, in the 0.5–4 second band using data recorded in July 2005 with a 19-station temporary broadband network. The source mechanism of the low-frequency events, obtained from the waveform inversion, included a volumetric component modeled as the resonance of a gently NNW-dipping, steam-filled crack located directly beneath the actively extruding part of the new dome and within 100 m of the crater floor, and a vertical single force attributed to the movement of the overlying dome.

Totally 68 studied microearthquakes demonstrated best-fit locations and source time functions indicating the repeated activation of the same source. In each of these mechanisms, the moment tensor was dominated by a volumetric component that was consistent with a northerly dipping crack, and  $F_z$  dominates the single-force components. Waite et al. (2008) considered that the source of low-frequency microearthquakes does not seem to be within the eruption conduit. The modeled crack could be along a preexisting zone of weakness that separates two older lava flows, for example. The most likely fluid in the crack was steam, owing to the abundant water, in the glacier and exsolved from the magma.

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## **RELEVANT WEBSITES**

BGVN, 1969–2009. Bulletin of Global Volcano Network. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.

GVP, 2011. Global Volcanic Program. Smithsonian Institution. <http://www.volcano.si.edu/world/volcano>.

## Chapter 18

# Acoustic Waves Generated by Volcanic Eruptions

“An earthquake in the air.” This was the first reaction of people watching the sudden and unanticipated jerk on the barograph records produced by the cataclysmic August 27, 1883 explosion of Krakatau volcano. The barograms were obtained from 16 observatories situated in various parts of Europe, from St. Petersburg in the east, to Valentia off the Irish coast. The barograph records allowed to see that the acoustic wave from this explosion had traveled around the Earth not once but seven times. This explosion was distinctly heard up to a distance of about 5000 km (Scott, 1883–1884; Winchester, 2003). Acoustic waves are generated also during small Strombolian and Vulcanian eruptions and phreatic eruptions or by simple degassing, during the movement of pyroclastic flows and rockfalls. Sustained acoustic signals resembling seismic tremors are also recorded.

In this chapter, the author presents the general characteristics of the acoustic waves observed during different types of volcanic activity together with the simultaneously recorded seismic waves. Making acoustic observations in conjunction with seismic observations at volcanoes allows one to separate shallow or subaerial processes from deeper (purely seismic) processes. The second advantage is that atmospheric acoustic waves travel with different propagation paths than seismic waves and can therefore provide constraints on source versus propagation effects. Another advantage is that infrasonic waves and acoustic-gravity waves can travel far from volcanoes and can be detected by remote stations.

### 18.1 INFRASONIC ACOUSTIC WAVES FROM SMALL VOLCANIC EXPLOSIONS (VEI 1 AND 2)

This type of acoustic wave is the most famous to volcanologists and is observed at many volcanoes. Infrasonic is the portion of the atmospheric acoustic wave spectrum below the low-frequency threshold of human hearing ( $\sim 20$  Hz; Beer, 1974). The commonest source of volcanic infrasonic is the atmospheric perturbation caused by the explosive outfluxes of volcanic volatiles (Johnson, 2003). Infrasonic pressure traces represent the time history of



atmospheric pressure perturbations relative to background atmospheric pressure. This excess pressure is usually very small compared to ambient atmospheric pressure ( $\sim 10^5$  Pa). Thus much volcanic infrasonic can be treated as linear elastic waves rather than nonlinear shock waves. Much volcanic infrasonic appears to be dominated by frequencies between 0.5 and 10 Hz. The infrasonic traces that are recorded several kilometers from the explosion source have peak excess pressures ranging from  $10^1$  to  $10^2$  Pa. An excess pressure of 10 Pa, which is the typical infrasonic amplitude recorded 1 km from a Strombolian source, corresponds to the sound pressure level of 115 dB. If this sound were in the audible frequency band, this level of sound would be heard as the noise equivalent of a pneumatic riveter (Johnson, 2003).

### 18.1.1 Waveforms and Spectra

Figure 18.1 demonstrates the simultaneous acoustic and seismic signals generated by the April 15, 1991 small explosion at Arenal volcano, Costa Rica and their Fourier spectra. The signals were recorded at the same site at a distance of 2.8 km from the crater. The acoustic signal represents a single impulsive compression with a peak frequency of 7.3 Hz contrasting with the peak frequency of the seismic signal of 1.1 Hz. The air wave arrival corresponding to this signal is seen on the seismic record also.

The infrasonic waveforms may be more complicated depending on the style of the explosive event (Johnson, 2003).

### 18.1.2 Families of Infrasonic Signals

Infrasonic signals, together with seismic signals from explosions, can form families of similar waveforms. Fig. 18.2 shows the family of seismic signals associated with the September 2003 degassing explosions at Shishaldin volcano (Petersen and McNutt, 2007). The acoustic signals accompanying these seismic signals are also characterized by similar waveforms. The seismic events have a mean cross-correlation coefficient of 0.8, while the acoustic events have a value of 0.9.

Cannata et al. (2009) applied the cross-correlation analysis to 987 infrasonic events recorded during the September 2007 to November 2007 Etna eruption, characterized by lava-fountaining episodes and degassing explosions. The events with correlation coefficients greater than 0.8 were grouped into 9 families which comprised about 95% of all considered events (Fig. 18.3). Tuning the cross-correlation coefficient threshold to 0.9, three clusters were recognized. Families 1–5 were grouped in cluster 1, families 6 and 7 in cluster 2, and families 8 and 9 in cluster 3. The infrasonic signals of the three clusters were not only similar in their waveforms but also grouped in terms of their quality factor and spectral peak frequency (Fig. 18.4). Cannata et al. (2009) considered that the differences between the waveform families and

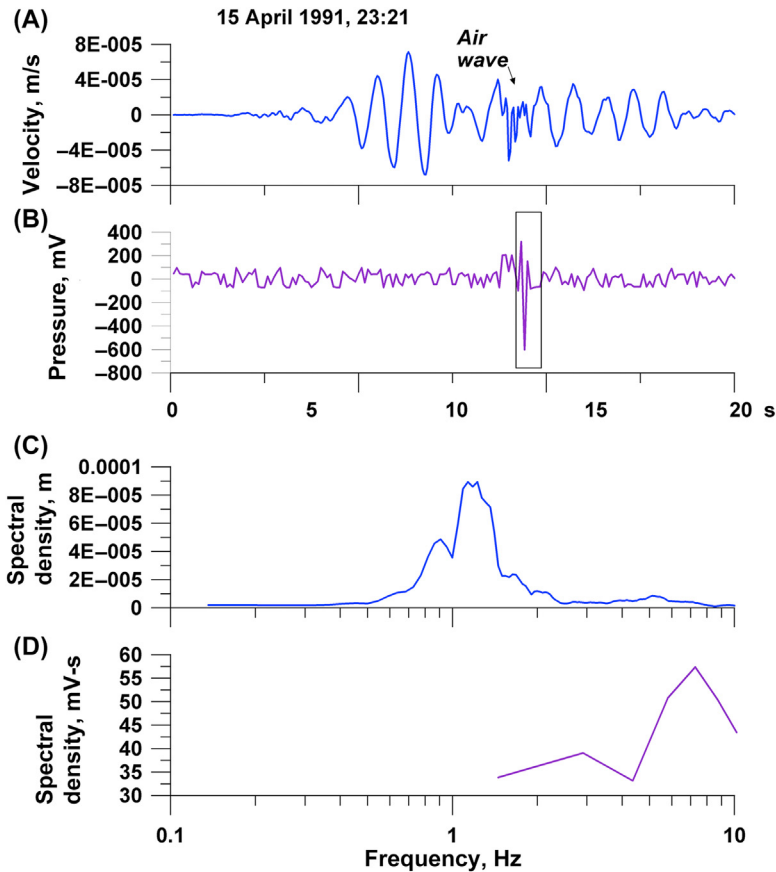


**FIGURE 18.1** Seismic (A) and infrasonic (B) signals generated by the April 15, 1991 explosion at Arenal volcano. The Fourier spectra are shown for the seismic signal (C) and the infrasonic signal (D). *Rectangle shows the infrasonic signal in (B).* (A) (B) Courtesy of Melson, W., Smithsonian Institution.

especially the clusters were linked to the different source vents and activities as also observed at other volcanoes (Ripepe and Marchetti, 2002).

The existence of waveform families and clusters suggests the repetitive excitation of stationary sources of infrasonic signals.

### 18.1.3 Source Location of the Infrasonic Events

The source of infrasonic waves may be approximated as a point source and its surface position can be very accurately located by examination of infrasonic arrivals which tend to be impulsive and possess self-similar waveforms across an array (Fig. 18.5). The acoustic propagation velocities in the atmosphere are

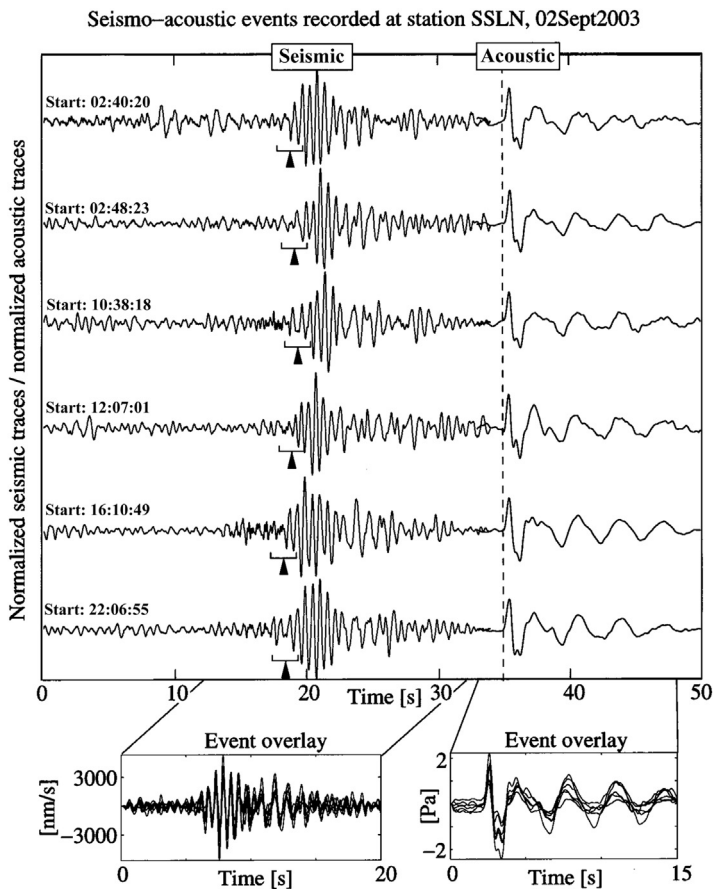


**FIGURE 18.2** Waveforms of similar seismic and corresponding similar acoustic signals produced during the explosive activity at Shishaldin volcano and recorded at a distance of 6.3 km from the crater. Seismic traces are high-pass filtered (above 0.8 Hz), and acoustic traces are bandpass filtered (0.02–3 Hz). The bottom two panels show an overlay of the six events. *From Petersen, T., McNutt, S.R. 2007. Seismo-acoustic signals associated with degassing explosions recorded at Shishaldin volcano, Alaska, 2003–04. Bulletin of Volcanology 69, 527–536.*

an order of magnitude slower than seismic velocities in the earth. For arrays, deployed at the epicentral distances within 5 km of the vent, explosion sources can be located with only a few meters error. Fig. 18.5 shows that the source of infrasonic waves generated by an explosion at the Karymsky volcano coincides with the crater of the volcano (Johnson et al., 2003).

The location of infrasonic events occurring in families may be done using the data of the stacked events belonging to the different families. In the case of the nine families of the Etna infrasonic events, described earlier (Cannata et al., 2009), the location of the infrasonic source was estimated by a

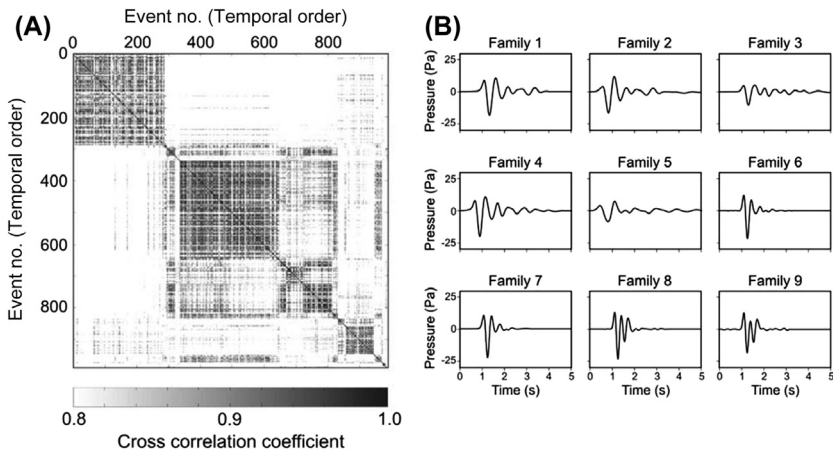


**FIGURE 18.3** Correlation matrix of the 987 infrasonic events (A) and the waveforms obtained by stacking the events belonging to nine families of infrasonic events, produced by the explosive activity at Etna volcano and recorded at a distance about 1.5 km from the active craters (B). From Cannata, A., Montalto, P., Privitera, E., Russo, G. 2009b. Characterization and location of infrasonic sources in active volcanoes: Mount Etna, September–November 2007. *Journal of Geophysical Research* 114, B08308. <https://doi.org/10.1029/2008JB006007>.

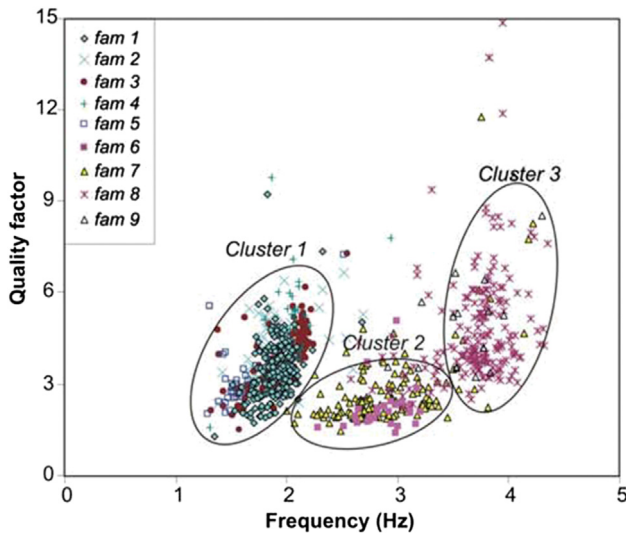


**FIGURE 18.4** Relationship between the peak frequency and the quality factor (calculated by Sompi method) for three clusters of infrasonic events of Etna formed by grouping of the nine families of infrasonic events shown in Fig. 17.3. From Cannata, A., Montalto, P., Privitera, E., Russo, G. 2009b. Characterization and location of infrasonic sources in active volcanoes: Mount Etna, September–November 2007. *Journal of Geophysical Research* 114, B08308. <https://doi.org/10.1029/2008JB006007>.

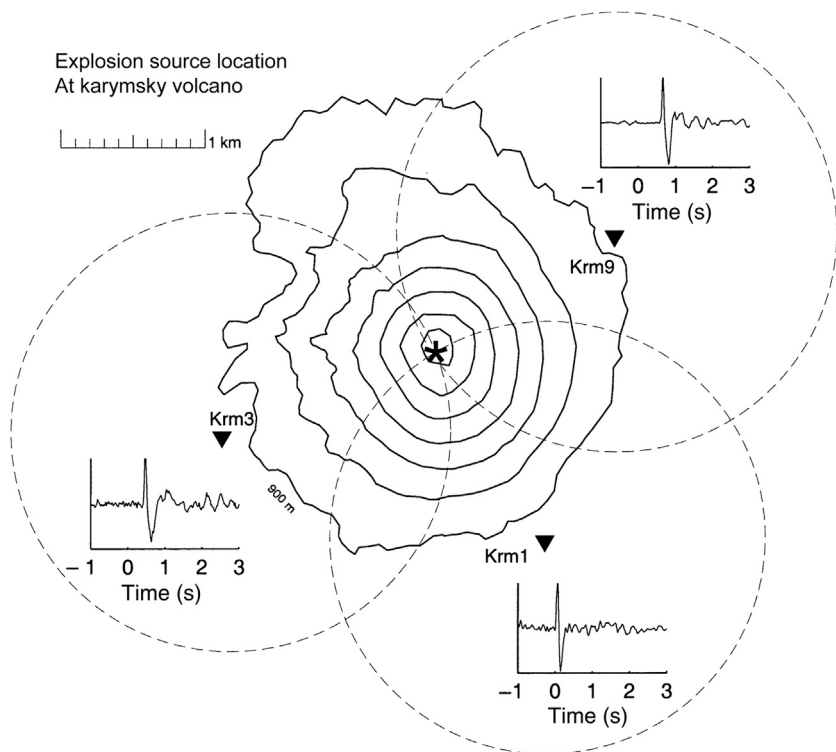


**FIGURE 18.5** Source location of acoustic signals determined from acoustic arrivals across an infrasonic array at Karymsky volcano. From Johnson, J.B. 2003. *Generation and propagation of infrasonic airwaves from volcanic explosions. Journal of Volcanology and Geothermal Research* 121, 1–14.

grid searching procedure over a surface of  $1.5 \text{ km} \times 1.5 \text{ km}$ , with space of 25 m, centered on the volcanic edifice. Infrasonic signals at the different stations were delayed and compared with the semblance function. The source was located in the node where the delayed signals showed the largest semblance value. As a result, the sources of the infrasonic events of all the nine families were associated with two Etna summit craters that were active from September 2007 to November 2007 (Fig. 18.6).

Among other methods of infrasonic source locating can be mentioned application of the semblance-based method (Ripepe and Marchetti, 2002) and the techniques based on the time lags between station pairs calculated by cross-correlation functions (Johnson, 2005).

All these direct methods located the surface projection of the infrasonic sources but did not allow for constraining the infrasonic source depth. To do it, the visual-seismic-infrasonic time delays could be very useful (e.g., Ripepe et al., 2001; Johnson, 2007; Yokoo et al., 2009). At the same time, the estimations of the depth of infrasonic sources, based on the time delay between simultaneous

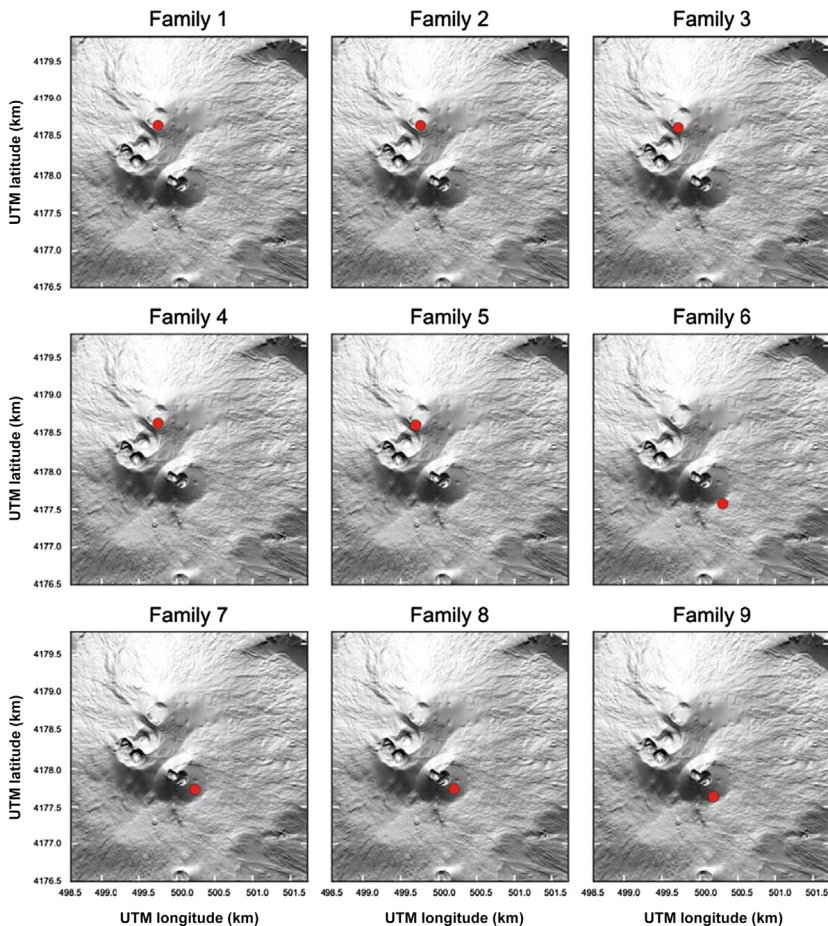


**FIGURE 18.6** Location of the stacked events of the nine families of infrasonic events of Etna. From Cannata, A., Montalto, P., Privitera, E., Russo, G. 2009b. Characterization and location of infrasonic sources in active volcanoes: Mount Etna, September–November 2007. *Journal of Geophysical Research* 114, B08308. <https://doi.org/10.1029/2008JB006007>.

seismic, acoustic, visual, and thermal records, strongly depend on the physical model of the sources chosen by authors and on uncertainties in the velocity models in air and solid within the near zone to the volcanic crater.

### 18.1.4 Relationship between the Amplitudes of the Seismic and Infrasonic Signals

Amplitudes of the seismic and infrasonic signals have a general tendency to be correlated but this relationship may depend on the style of explosive event and its surface manifestations. Fig. 18.7 shows the plots of the seismic and acoustic

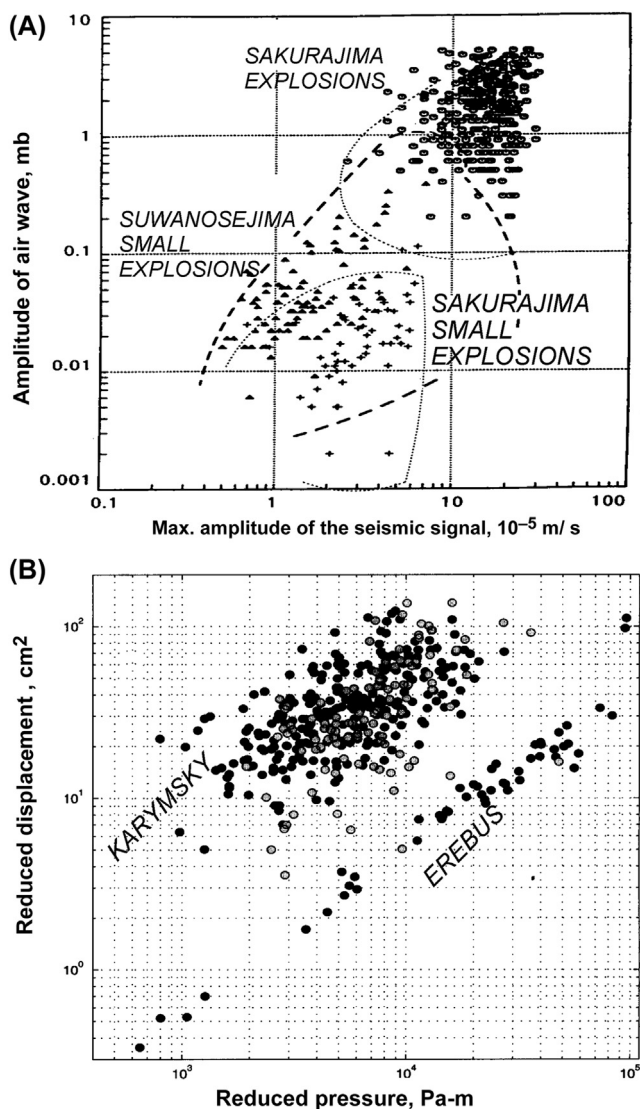


**FIGURE 18.7** Relationships between the amplitudes of infrasonic and seismic waves that were produced by explosion earthquakes at Sakurajima and Suwanosejima volcanoes (A) and Karymsky and Erebus volcanoes (B). In (B), 427 Karymsky 1998 explosions are shown with the gray symbols, 189 Karymsky 1999 explosions with the open symbols, and 64 Erebus 1999–2000 explosions are shown with the black symbols. Compiled from Iguchi, M., Ishihara, K. 1990. Comparison of earthquakes and air-shocks accompanied with explosive eruptions at Sakurajima and Suwanosejima volcanoes. *Annals Disaster Prevention Research Institute Kyoto University* 33B-1, 1–12 and Johnson, J.B., Aster, R.C., Ruiz, M.C., Malone, S.D., McChesney, P.J., Lees, J.M., Kyle, P.R. 2003. Interpretation and utility of infrasonic records from erupting volcanoes. *Journal of Volcanology and Geothermal Research* 121, 15–63.



parameters for the explosive events observed at Sakurajima and Suwanosejima volcanoes (Iguchi and Ishihara, 1990) and at Karymsky and Erebus volcanoes (Johnson et al., 2003). The explosive events observed at Sakurajima and Suwanosejima volcanoes (Fig. 18.7A) demonstrate the general correlation between the maximum amplitude of the compressional phase of infrasonic waves and the maximum amplitude of the seismic signals. Iguchi and Ishihara (1990) showed that this relationship depends on the style of an eruption. They compare the acoustic and seismic signals produced by large explosions at Sakurajima and Suwanosejima and small explosions at Sakurajima. Within the same amplitude range of the seismic signals ( $(5-7) \times 10^{-5}$  m/s), the amplitudes of infrasonic waves produced by the explosions at Suwanosejima are larger than those produced by small eruptions at Sakurajima but smaller than those produced by the large Sakurajima explosions.

Figure 18.7B displays a relatively fixed relationship between seismic reduced displacement and acoustic reduced pressure for the Erebus explosion events. However, the intensity of Karymsky infrasonic, shown in the figure, appears to have little correlation with seismic amplitude. Johnson et al. (2003) attribute this variability in energy partitioning to changeable source locations within the conduit and/or muffling of the infrasonic pulse due to an overlying plug of material. They believe that degassed andesite choking the upper portion of the Karymsky conduit possesses a considerably higher viscosity than the Erebus phonolitic lava lake. Later, Johnson and Aster (2005) estimated the volcano acoustic–seismic ratio during Strombolian eruptions (which is the ratio of elastic energy propagated through the atmosphere and into the earth) as equal to 8 for Erebus eruptions and equal to 0.2–1.5 for Karymsky eruptions.

Seismoacoustic study of the explosions that occurred at the andesitic Tokachi-dake volcano, Hokkaido (Okada et al., 1990) showed that the amplitudes of the seismic signals may not correlate with the amplitudes of the air waves (Fig. 18.8) when the surface manifestations of explosive events are different. Okada et al. (1990) recognized that the explosions accompanied by pyroclastic flows were characterized by larger seismic amplitudes with a similar intensity of the air waves.

## 18.2 LONG-PERIOD ACOUSTIC AND ACOUSTIC-GRAVITY WAVES FROM LARGE VOLCANIC EXPLOSIONS (VEI 4–6)

Large volcanic eruptions, which form high eruption column with a great heat injection into the atmosphere, generate large-scale atmospheric pressure perturbations as long-period acoustic waves with periods greater than 270 seconds or acoustic-gravity waves that are driven by buoyancy forces and propagate in a gravitational field of atmosphere. These long-period waves are characterized by low attenuation and may propagate for thousands of kilometers traveling around the Earth (Beer, 1974; Johnson, 2003).



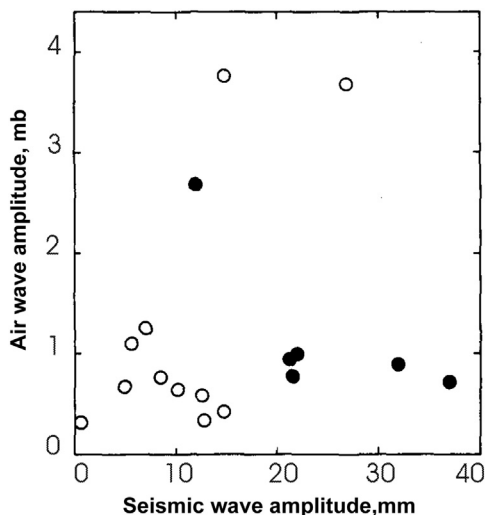


**FIGURE 18.8** Relationship between the amplitudes of air waves and seismic waves that were produced by explosion earthquakes at Tokachi-dake volcano, Hokkaido during 1988–89. The amplitude of seismic waves is peak-to-peak maximum amplitude recorded by a short-period seismograph at a distance of about 7 km from the crater; the amplitude of the air wave was measured at a distance of about 3 km from the crater. The *filled circles* indicate the explosions accompanied by pyroclastic flows. From Okada, H., Nishimura, Y., Miyamachi, H., Mori H., Ishihara, K. 1990. Geophysical significance of the 1988–1989 explosive eruptions of Mt. Tokachi, Hokkaido, Japan. *Bulletin of the Volcanological Society of Japan* 35, 175–203.

### 18.2.1 Near-Field Waveforms of the Long-Period Acoustic Waves

Figure 18.9 shows the long-period 3-hour records of the atmospheric pressure perturbations associated with four great Plinian and sub-Plinian (VEI 4–5; Newhall and Self, 1982) explosions at volcanoes Mount Pelée (1902), Bezymianny (1956), Sheveluch (1964), and Mount St. Helens (1980) collected by Firstov (2007). These waveforms were recorded at distances from 22 to 54 km from the crater. The maximum period of these records ranges between 580 seconds for Bezymianny volcano and 1200 seconds for Sheveluch volcano, and the range of maximum double amplitudes of pressure is between 350 Pa for Mount Pelée and 2100 Pa for Bezymianny volcano.

Audible sounds were also reported in association with these great explosions. The 1902 Mount Pelée explosion noted “A dreadful scream, then another much weaker groan, like a stifled death rattle” (Zebrowski, 2002). Gorshkov and Bogoyavlenskaya (1965) reported that the acoustic signal generated by the March 30, 1956 gigantic explosion of Bezymianny volcano was not heard at near distances from the volcano, at least within 45 km from the crater; only a weak boom was reported. During the 1980 Mount St. Helens great explosion, it was noted that the blast seemed to have been surprisingly quiet in the vicinity of the mount. At the same time, loud noises

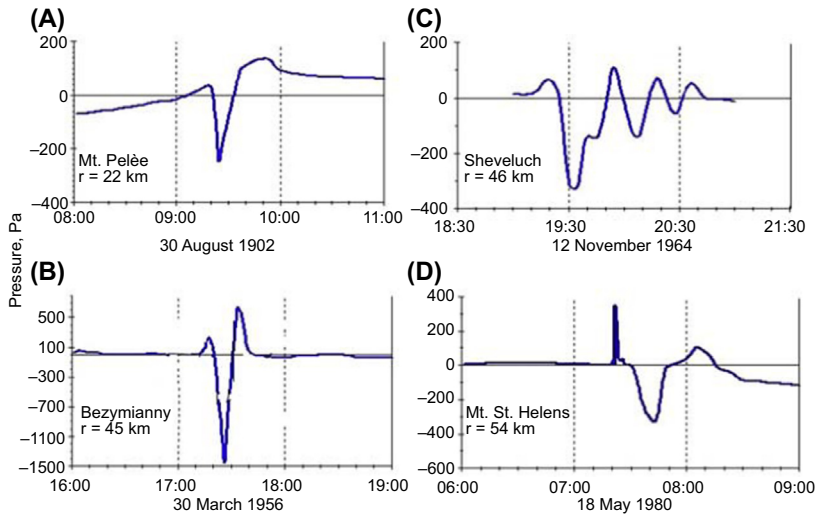


**FIGURE 18.9** Long-period 3-h records of the atmospheric pressure perturbations associated with four great Plinian explosions at volcanoes Mount Pelée (A), Bezymianny (B), Sheveluch (C), and Mount St. Helens (D). From Firstov, P.P. 2007. Long-period acoustic waves as a source of information about the large explosive eruptions. *Proc. Annual Conf. of Day of Volcanologist. Institute of Volcanology and Seismology, Petropavlovsk-Kamchatsky*, pp. 83–94 (in Russian).

due to the eruption were heard hundreds of kilometers away (Rosenbaum and Waite, 1981). A similar phenomenon was reported for the 1883 climactic eruption of Krakatoa which was heard 5000 km away but not on neighboring islands (Symons, 1888).

Firstov (2007) noted that the waveforms for the explosions of Mount Pelée and Bezymianny volcanoes, shown in Fig. 18.9, were similar and were characterized by two compressive and one rarefaction phases. The acoustic records of the Sheveluch and Mount St. Helens explosions were more complicated. Tokarev (1967) considers that the complicated acoustic waveform of the Sheveluch explosion was produced by not a single explosion but a sequence of explosive events.

### 18.2.2 Far-Field Registrations of the Long-Period Acoustic Waves

The barograph records allow to see that the acoustic waves from the explosions travel around the Earth not once but several times. As was mentioned previously, the acoustic waves from the Krakatoa explosion traveled around the Earth seven times. The acoustic waves from the Bezymianny explosion traveled around the Earth 1.5 times (Passechnik, 1958).

### 18.2.2.1 Far-Field Signals of Acoustic Waves

Figure 18.10 shows the copies of the microbarographic records of the long-period acoustic waves produced by the 1980 explosion at Mount St. Helens volcano obtained at the acoustic stations of North America (distances from 927 to 3950 km) and Japan (distances from 6970 to 8240 km). There are seen two impulses in the beginning of all records. The duration of them is about 5–7 minutes. Mikumo and Bolt (1985) considered that these two impulses are related to the initial stage of the paroxysmal eruption. The first was attributed to the generation of the rockslide avalanche and the northward horizontal blast of ash and debris and the second was attributed to a vertical Plinian eruption column.

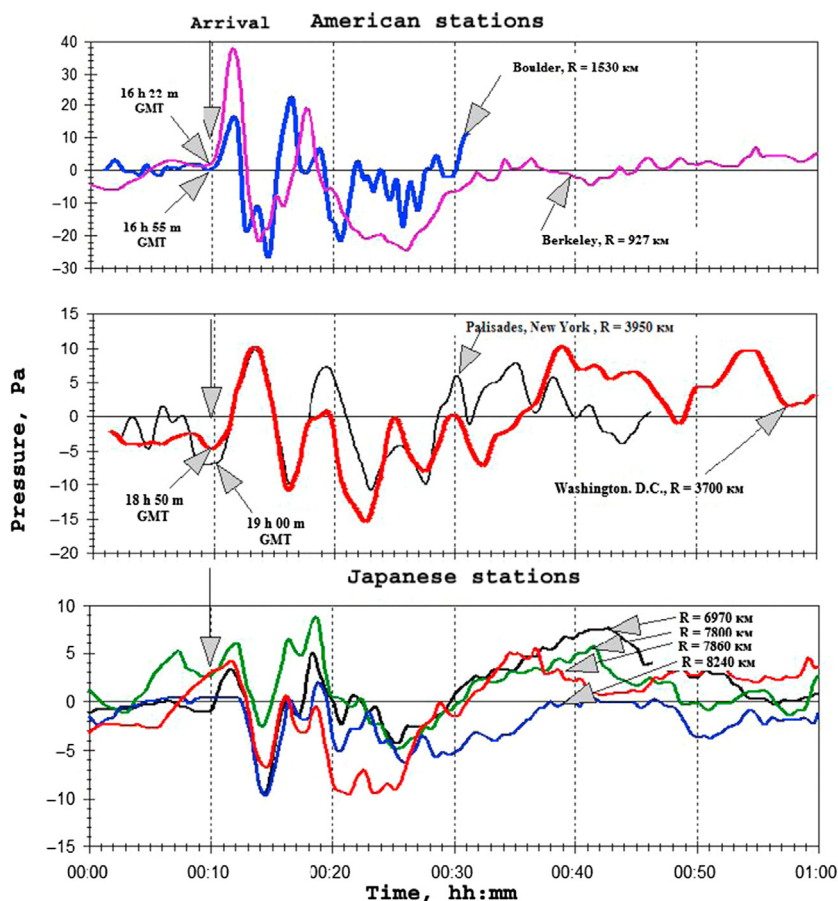


**FIGURE 18.10** Long-period far-field acoustic perturbations associated with the 1980 explosion at Mount St. Helens volcano and recorded at the acoustic stations of North America (Reed, 1987) and Japan (Sawada et al., 1982). From Firstov, P.P. 2007. Long-period acoustic waves as a source of information about the large explosive eruptions. *Proc. Annual Conf. of Day of Volcanologist. Institute of Volcanology and Seismology, Petropavlovsk-Kamchatsky*, pp. 83–94 (in Russian).

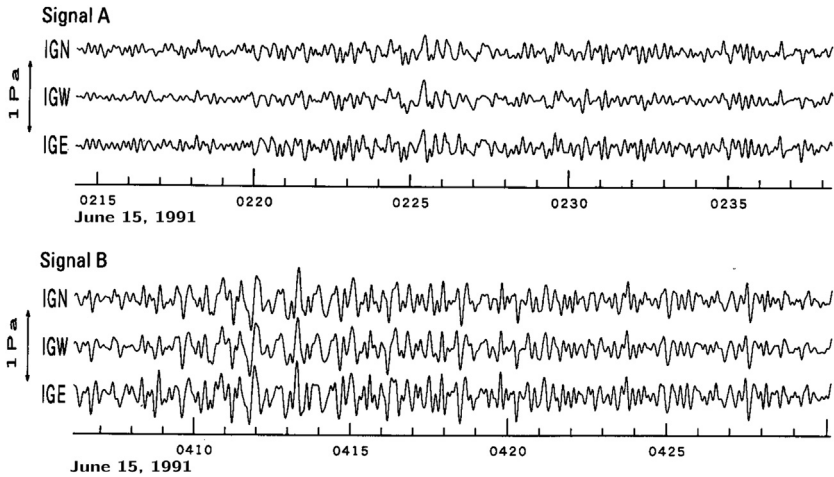


**FIGURE 18.11** Acoustic signals recorded at Kariya Observatory from the explosions of Pinatubo volcano on June 15, 1991. The three traces in each panel show the output of three microphones of the tripartite array. From Tahira, M., Nomura, M., Sawada, Y., Kamo, K. 1996. *Infrasonic and acoustic-gravity waves generated by the Mount Pinatubo eruption of June 15, 1991*. In: Newhall, C.G., R.S. Punongbayan (Eds), *Fire and Mud*. University of Washington Press, Seattle, pp. 601–614.

Tahira et al. (1996) describe the far-field signals of acoustic waves produced by the June 15, 1991 Pinatubo eruption. They discuss the records obtained at Kariya Observatory, Japan, 2770 km to the NE of the volcano. Waves that arrived between 02:45 and 02:54 after major explosions are interpreted as the waves that propagated along the shorter great circular path in the atmosphere. The strong acoustic signal (the parts of it are shown in Fig. 18.11) lasted for almost 10 hours. The signals propagated with a velocity between 265 and 280 m/s; their maximum amplitudes varied between 0.17 and 1.73 Pa.

#### 18.2.2.2 Far-Field Signals of Acoustic-Gravity Waves

Acoustic-gravity waves are waves, exhibiting both buoyant and compressional motion, typically with millihertz frequencies and long wavelengths (tens of kilometers). They are represented by long-period Lamb-type waves, propagating along Earth's surface, with group velocities near the mean sound speed of the lower atmosphere. Lamb waves are associated with the largest atmospheric explosions from volcanic eruptions and have periods on the order of several to hundreds of minutes (Matoza et al., 2022).

A long train of acoustic-gravity waves produced by the June 15, 1991 Pinatubo eruption and associated with extremely large movements of air parcels was recorded at five microbarographic stations in Japan (Tahira et al., 1996).

The high-pass filtered waveforms and their power spectra are shown in Fig. 18.12. The group velocity of the leading part of the wave is 354 m/s. The duration of the acoustic-gravity waves is 03:50, which is comparable to the duration of the strong infrasonic signals at Kariya shown in Fig. 18.11. The spectral peaks obtained from the power spectra were noticed at the frequencies of 1.2, 1.9, 2.4, and 4.4 mHz, respectively; the longest period is 13.9 minutes.

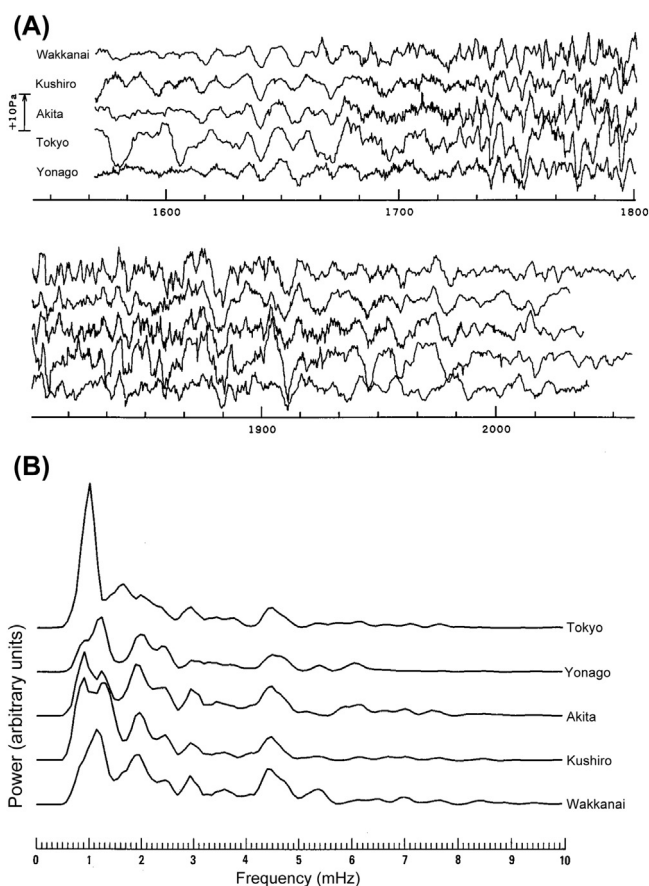


**FIGURE 18.12** Acoustic-gravity waveforms from the explosions of Pinatubo volcano, observed at five stations in Japan on June 15, 1991 (A), and their power spectra (B). The data were high-pass filtered to eliminate the incoherent noise with periods longer than 34 min and retain periods shorter than 17 min. From Tahira, M., Nomura, M., Sawada, Y., Kamo, K. 1996. *Infrasonic and acoustic-gravity waves generated by the Mount Pinatubo eruption of June 15, 1991*. In: Newhall, C.G., R.S. Punongbayan (Eds), *Fire and Mud*. University of Washington Press, Seattle, pp. 601–614.

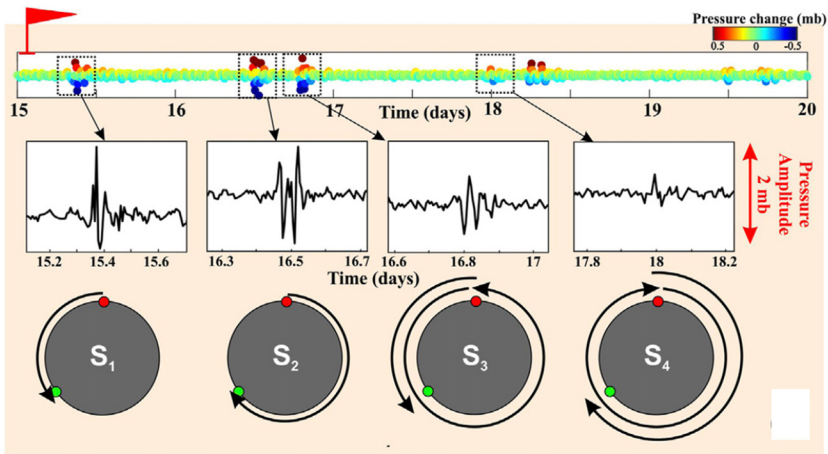


**FIGURE 18.13** Atmospheric pressure disturbance generated by the 2022 Hunga Tonga volcanic eruption (Dalal et al., 2023). Atmospheric pressure fluctuation data from Kahului, Hawaii meteorological station, showing the four pressure fluctuations ( $S_1$ – $S_4$ ) from 15–19 January are presented in the upper panel. Enlarged versions of four individual signals are shown in the middle panel. Lowest panel shows the representative travel path of the Lamb wave around the globe. From Dalal, P., Kundu, B., Panda, J. and Jin, S., 2023. Atmospheric Lamb wave pulse and volcanic explosivity index following the 2022 Hunga Tonga (South Pacific) eruption. *Frontiers Earth Sciences*, 10, 931545. <https://doi.org/10.3389/feart.2022.931545>.

The Plinian (VEI 6; Poli and Shapiro, 2022) explosion of 15 January 2022, produced by the submarine volcano Hunga Tonga-Hunga Ha’apai (HTHH), generated a volcanic plume of more than 58 km high (Yuen et al., 2022). A broad range of atmospheric waves were observed globally by various ground-based and spaceborne instrumentation networks (Fig. 18.13). This was one of the most remarkable phenomena, which was detected as fluctuations of atmospheric pressure at numerous observation sites while circling the Earth for a few days after this historic volcanic event occurred. The acoustic-gravity surface-guided Lamb wave ( $f \leq 0.01$  Hz) was the most prominent of the atmospheric waves produced by the Plinian explosion (Matoza et al., 2022).

Fig. 18.14 (A,B) demonstrates the infrasound and barometric waveforms, recorded at the Italian stations, situated at Campi Flegrei (CF) and Etna, respectively. Both barometric signals indicate that the frequency content and the arrival time, corresponding to an average velocity propagation of about 310 m/s from HTHH, are characteristic for the Lamb wave. With a delay of about 1 h, the infrasonic signals mark the arrival of higher-frequency waves that fall into the sensor frequency band (above about 0.2 Hz). The applied Continuous Wavelet Transform (Fig. 18.14C) shows a clear dispersion line due to the lower propagation velocities for higher frequencies (Madonia et al., 2023).

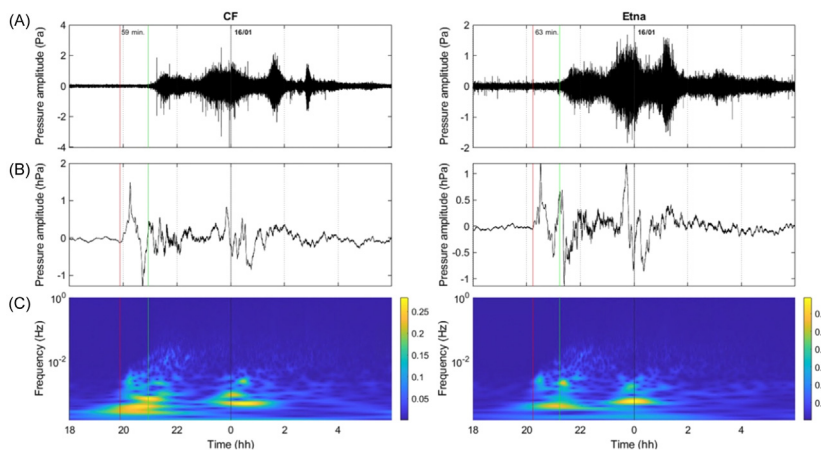


**FIGURE 18.14** (A) Infrasound waveforms recorded at Campi Flegrei (CF) and Etna CBAC stations showing the acoustic wave arrival; the signal is filtered with a low-pass filter with 10 Hz cutoff. (B) Barometric waveforms, recorded at the same stations, showing the Lamb wave arrival; the signal is filtered with a high-pass filter with 0.093 mHz (3 h period) cutoff. (C) Continuous wavelet transforms of the barometric data show a dispersion line due to lower propagation velocities for higher frequency. The vertical red and green lines indicate the Lamb wave and the acoustic wave arrivals, respectively. In the figure, the delay times (59 and 63 min for CF and Etna, respectively) between Lamb and acoustic wave arrivals are indicated. *From Madonia, P., Bonaccorso, A., Bonforte, A., et al, 2023. Propagation of Perturbations in the Lower and Upper Atmosphere over the Central Mediterranean, Driven by the 15 January 2022 Hunga Tonga-Hunga Ha'apai Volcano Explosion. Atmosphere, 14, 65. <https://doi.org/10.3390/atmos14010065>.*

Takamori et al. (2023) published the records of atmospheric pressure and horizontal strain of the first Lamb wave ( $L_1$ ), obtained at an underground observatory in Kamioka, Japan using the laser strainmeter with a 1500-m baseline length (GIF). Fig. 18.15 shows the detided strain observed by the GIF at the arrival of the  $L_1$  wave that travelled along the minor arc from the south to arrive at approximately 11:30 UTC, i.e., 7 h and 15 min after the eruption. A clear negative correlation between the pressure and strain can be seen immediately (Fig. 18.15). The maximum variations of the atmospheric pressure and strain were approximately 1.9 hPa and  $-4.0 \times 10^{-10}$ , respectively, where the negative sign means compression.

Over 6 days, there was observed global propagation of at least four minor-arc Lamb wave passages and three major-arc (antipodal) passages (Matoza et al., 2022). The number of Lamb wave passages, observed for the HTHH event (four, plus three antipodal), was approximately the same as observed for the 1883 Krakatau eruption. Comparison of the digitized and calibrated waveforms for the first two arrivals (Matoza et al., 2022) demonstrated the similarity of the acoustic-gravity Lamb waves for both explosive



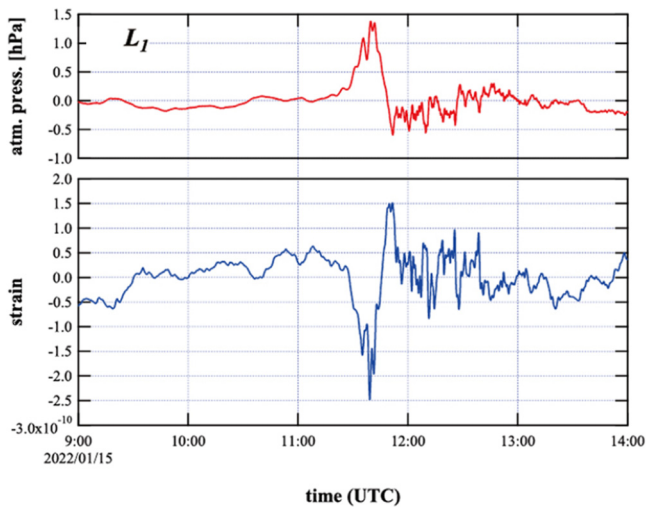


**FIGURE 18.15** Variations of the detided strain and atmospheric pressure during the arrival of the first direct Lamb wave  $L_1$ , induced by the 2022 Hunga Tonga-Hunga Ha’apai volcanic eruption, in Kamioka, Japan. The records were realized by a 1500-m laser strainmeter. From Takamori, A., Araya, A., Miyo, K. et al., 2023. Ground strains induced by the 2022 Hunga-Tonga volcanic eruption, observed by a 1500-m laser strainmeter at Kamioka, Japan. *Earth Planets Space* 75, 98. <https://doi.org/10.1186/s40623-023-01857-w>.

events. Measurements of the Lamb wave peak-to-peak pressure amplitudes, generated by the HTHH and Krakatau events and recorded at the global seismic network, indicated that they were comparable in size. The same time, they were more than an order of magnitude larger than those, generated by the VEI 5 1980 Mount St. Helens eruption (Matoza et al., 2022; Reed, 1987; Symons, 1888).

### 18.3 ACOUSTIC WAVES PRODUCED BY THE LAVA DOME COLLAPSE AND THE PROPAGATION OF PYROCLASTIC FLOW AND ROCKFALLS

#### 18.3.1 Dome Collapse

Yamasato (1997) carried out the studies of the acoustic waves produced by pyroclastic flows that followed the dome collapse at Unzen volcano. Fig. 18.16 shows the simultaneous seismic and infrasonic records obtained during the dome collapse occurring on October 18, 1992. A small impulsive infrasonic signal was excited at the time of the dome collapse and larger infrasonic waves were excited when the lava blocks fell onto the slope.



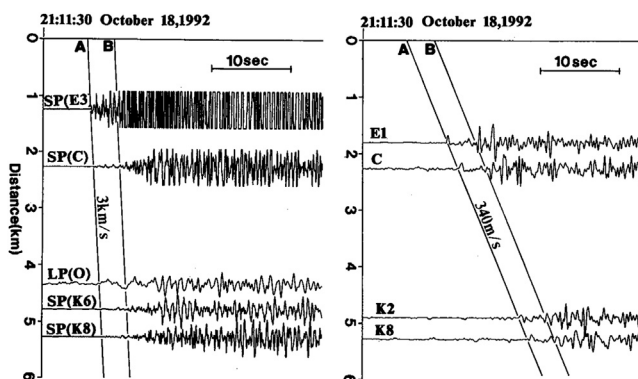


**FIGURE 18.16** Example of simultaneous seismic (left) and infrasonic (right) signals from a dome collapse at Unzen volcano. Seismic records were obtained by short-period (SP) and long-period (LP) vertical seismographs. Infrasonic records were obtained by low-frequency microphones. The phases A and B are assumed to correspond to the start of dome collapse and the fall of lava blocks onto the slope, respectively. From Yamasato, H. 1997. *Quantitative analysis of pyroclastic flows using infrasonic and seismic data at Unzen volcano, Japan. Journal of Physics of the Earth* 45, 397–416.

Oshima and Maekawa (2001) also performed similar observations at Unzen volcano during the October 1992 pyroclastic flow activity. Their instruments were deployed at a distance of 4 km from the new lava dome. They revealed from the correlation of infrasonic waves with video images that during the collapse of lava blocks falling from the dome, which generated a pyroclastic flow, the successive impulsive waves and the low-frequency wave ( $<0.8$  Hz) were excited (Fig. 18.17). Oshima and Maekawa (2001) supposed that the impulsive wave was excited by an explosive blow-out of the pore gases which were released by the breakdown of the falling blocks and the low-frequency wave was excited by the volumetric change which resulted from the expansion of the pore gases released by pulverization of the falling blocks.

## 18.3.2 Pyroclastic Flow Propagation

### 18.3.2.1 Observations in Near Zone

Figure 18.18 shows the simultaneous seismic and infrasonic records obtained during the pyroclastic flow propagation at Unzen volcano occurring on May 5, 1993 at a distance of 1–6 km from the crater (Yamasato, 1997). It is seen that the amplitude variation of the infrasonic signal recorded at these distances was similar to that of the seismic signal, and the infrasonic signals became attenuated when the seismic waves decayed.

Simultaneous seismic and infrasonic records, obtained at a distance of 16 km from the crater during the October 13, 1984 pyroclastic flow

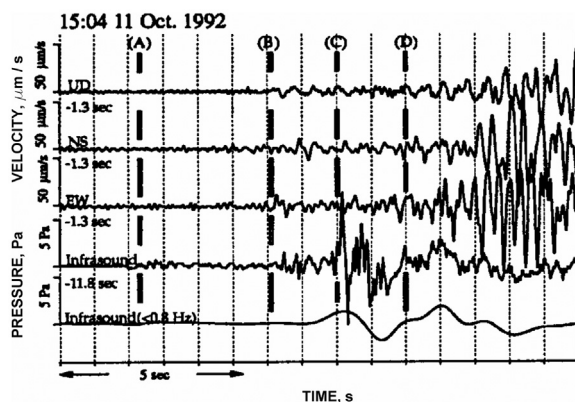


**FIGURE 18.17** Comparison of infrasonic and seismic signals with visible events for the lava collapse that generated pyroclastic flow at Unzen volcano on October 11, 1992, at 15:04. The lowest trace indicates filtered infrasonic waves with low-pass Butterworth filter. Thick broken lines (A–D) correspond to visible events: (A) a lava block was detached from the dome, (B) the basal part of the falling block landed on the slope, (C) the block violently crumbled from below and explosively discharged the gas-ash-debris complex on the slope, and (D) the block broken into pieces and the complex started to rush down. *From Oshima, H., and Maekawa, T. 2001. Excitation process of infrasonic waves associated with Merapi-type pyroclastic flow as revealed by a new recording system. Geophysical Research Letters 28, 1099–1102.*

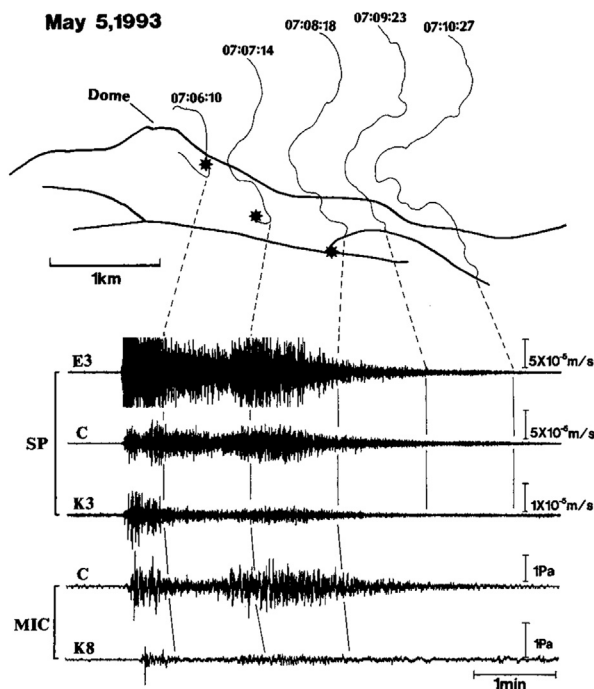


**FIGURE 18.18** Illustration of an advancing pyroclastic flow at Unzen volcano and the corresponding seismic (SP) and infrasonic (MIC) signals. The stars indicate the source locations of the infrasonic signals. *From Yamasato, H. 1997. Quantitative analysis of pyroclastic flows using infrasonic and seismic data at Unzen volcano, Japan. Journal of Physics of the Earth 45, 397–416.*

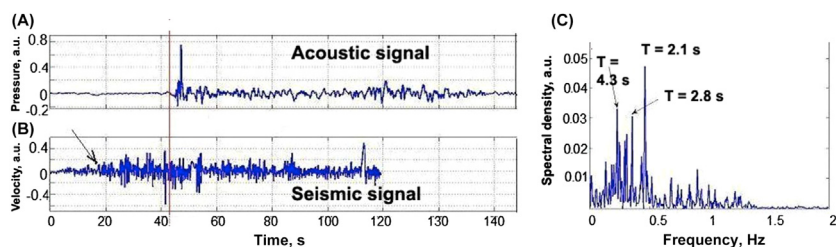


**FIGURE 18.19** Infrasonic (A) and seismic (B) records, obtained during the October 13, 1984 pyroclastic flow movement at Bezymianny volcano at a distance of 16 km from the crater and the Fourier spectrum of infrasonic record (C). Arrow shows the beginning of the seismic signal. Vertical line shows the beginning of the infrasonic signal. From Firstov, P.P., Tristanov A.B. 2010. Generation of acoustic waves by the explosive eruptions of andesitic volcanoes and associated pyroclastic flows. *Proc. Annual Conf. of Day of Volcanologist, Institute of Volcanology and Seismology, Petropavlovsk-Kamchatsky*. pp. 50–61 (in Russian).

propagation at Bezymianny volcano (Firstov and Tristanov, 2010), show the impulsive acoustic signal that arrived in 35 seconds after the beginning of the seismic signal (Fig. 18.19). No similarity in the amplitude variation of the infrasonic and seismic signals recorded at this distance was observed. The spectral peaks for the acoustic signal were at the intervals of 2.1, 2.8, and 4.3 seconds. Firstov and Tristanov (2010) reported that the generation of intense infrasound signals began in about 25 seconds after entering the pyroclastic flow from the steep narrow trough into the gently sloping wide ravine.

An infrasonic array, installed at 3.5 km from Soufriere Hills Volcano, has successfully detected the infrasonic waves associated with large Vulcanian explosions and the propagation of energetic pyroclastic fronts during May 2008, July 2008, and December 2008 (Ripepe et al., 2010). The array recorded clear signals with durations of 400–1500 seconds (Fig. 18.20). Coherent infrasonic signal across the array started with an impulsive onset and reached peak amplitudes of 40–80 Pa during the most vigorous phase of the joint explosion-pyroclastic flow event. It is difficult to separate the acoustic signals produced by the explosion and by the pyroclastic flow on these records, obtained near the volcanic crater. Only for the December 3 event, the signal between 1500 and 2500 seconds is large enough to be identified as a record of pyroclastic flow. The infrasound, recorded at the array for this explosion-pyroclastic flow event, was relatively broadband, ranging from 0.4 to 7 Hz.

### 18.3.2.2 Distant Observations

Firstov and Tristanov (2010) presented the observations of the acoustic signals produced by the May 9, 2006 eruption at Bezymianny volcano when the propagation of pyroclastic flow generated the co-ignimbrite plume that reached the height of 11 km above sea level (a.s.l.; Fig. 18.21A and B). These signals were recorded at two acoustic stations situated at the distances of 340 and 361 km from the volcano. For comparison, the seismic signal obtained at a distance of

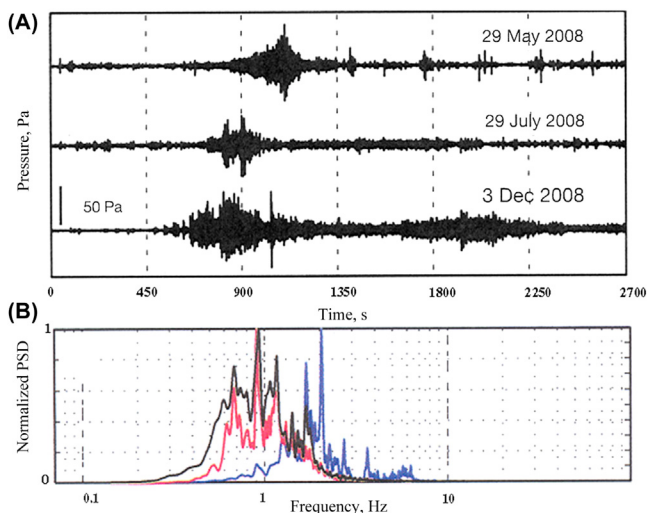


**FIGURE 18.20** Infrasonic associated with Vulcanian eruptions and associated pyroclastic flows on May 29, 2008; July 29, 2008; and December 3, 2008 recorded by the infrasonic array installed at 3.5 km from the Soufrière Hills volcano (A) and Fourier spectra of the infrasonic waveforms (B). Note the July 29 (blue) and December 3 (black) spectra are peaked at 0.9 Hz, while the smaller event on May 29 (red) shows the largest energy at 2 Hz. From Ripepe, M., De Angelis, S., Lacanna, G., Voight, B. 2010. Observation of infrasonic and gravity waves at Soufrière Hills Volcano, Montserrat. *Geophysical Research Letters* 37, L00E14 with modifications.

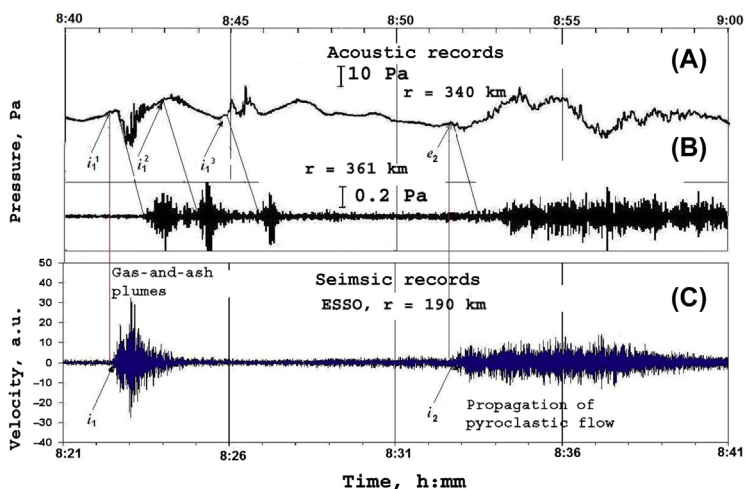


**FIGURE 18.21** Acoustic [(A), broadband channel, 0.003–0.3 Hz; (B), bandpass filtered record at 0.4–8.0 Hz] and seismic [(C), short-period velocity channel] records of the May 9, 2006 explosive event at Bezmyanny volcano and associated pyroclastic flow. The phases associated with the gas-and-ash plume activity are marked with, the arrivals of the pyroclastic flow phase are marked with  $i_2$  at the seismic signal and with  $e_2$  at the infrasonic signal. From Firstov, P.P., Tristano, A.B. 2010. Generation of acoustic waves by the explosive eruptions of andesitic volcanoes and associated pyroclastic flows. *Proc. Annual Conf. of Day of Volcanologist, Institute of Volcanology and Seismology, Petropavlovsk-Kamchatsky*. pp. 50–61 (in Russian).

190 km from the volcano is shown also (Fig. 18.21C). Two phases of the eruption, representing the explosive activity and the propagation of pyroclastic flow (see Section 13.2.3), are well indicated on the infrasound and seismic records. Infrasound record allows to discriminate at least three phases (, and) within the activity of explosive gas-and-ash plumes.

### 18.3.3 Large Rockfall Propagation

Very-long-period (VLP) infrasonic signals (Fig. 18.22) were produced by an unusually large rockfall that occurred on May 29, 2006, at Mount St. Helens

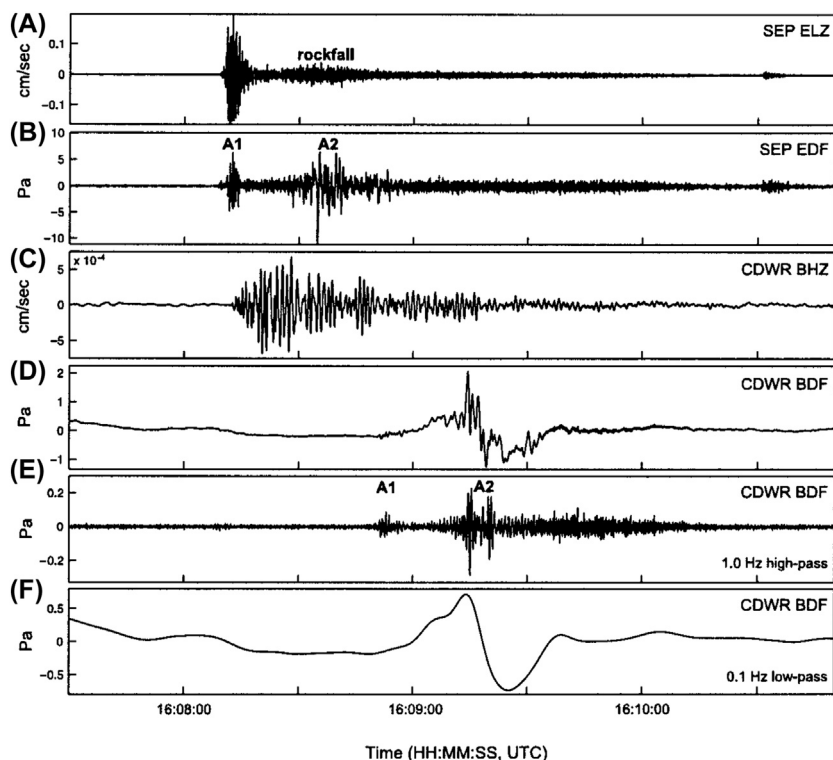


**FIGURE 18.22** Seismic and infrasonic signals associated with large rockfall that occurred on May 29, 2006, at Mount St. Helens. (A) Low-gain vertical component record from station SEP (0.4 km from the vent). (B) Infrasonic signals recorded by a microphone at station SEP. A1 and A2 labels indicate the timing of notable high-frequency infrasonic signals. (C) Broadband vertical component record from station CDWR (13.4 km from the vent). (D) Unfiltered infrasonic signals recorded by a microphone at station CDWR. (E) Same as (D), but filtered with a 1.0 Hz high-pass filter. A1 and A2 labels indicate the timing of notable high-frequency infrasonic signals. (F) Same as (D), but filtered with a 0.1 Hz low-pass filter. From Moran, S.C., Matoza, R.S., Garces, M.A., Hedlin, M.A.H., Bowers, D., Scott, W.E., Sherrod, D.R., Vallance, J.W. 2008c. Seismic and acoustic recordings of an unusually large rockfall at Mount St. Helens, Washington. *Geophysical Research Letters* 35, L19302.

(Moran et al., 2008a,b). The field observations showed that no explosive activity or lava dome collapse was observed before this event. However, an atmospheric plume had risen to 6000 m a.s.l. and the seismic signal produced by this event was equivalent to the  $M$  3.1 earthquake (Fig. 18.22A and C).

Two large amplitude, high-frequency infrasonic pulses, marked as A1 and A2 in Fig. 18.22, were recorded. Accompanying the higher frequency infrasonic signals was a  $\approx 50$ -s-period VLP infrasonic pulse that can be seen in the raw and low-pass filtered versions of the microbarometer records obtained at station CDWR at a distance of 13.4 km from the vent (Fig. 18.22D and F). The onset of the VLP signal is coincident with the arrival of the A1 pulse, suggesting that it is related with the initial collapse of the rock mass. VLP pulse featured an initial pressure increase followed by a pressure decrease before returning to zero (Fig. 18.22F).

## 18.4 ACOUSTIC WAVES PRODUCED DURING VOLCANIC MICROEARTHQUAKE SWARMS (“DRUMBEATS”)

Seismicity during the 2004–08 eruption of Mount St. Helens has been characterized by a sustained sequence of microearthquakes occurring beneath the dome and named “drumbeats” because of their highly repetitive and regular nature and constant delay time between successive events (see Chapter 17). Matoza et al. (2007) demonstrated that the majority of these seismic events had intermittent infrasonic twins, the infrasonic arrivals with waveforms mimicking the seismic event waveforms and arriving after the seismic event with a velocity and time delay appropriate for an acoustic wave (Fig. 18.23). Matoza et al. (2009a) showed that the acoustic signals were generated by rapid degassing associated with a fluid source of microearthquakes.

## 18.5 UTILITY OF THE ACOUSTIC SIGNALS FOR VOLCANO ACTIVITY MONITORING

### 18.5.1 Estimation of the Energy of Eruptive Events

#### 18.5.1.1 Large Explosions

The source energy from the microbarographic records was estimated for several large volcanic eruptions. For the 1980 Mount St. Helens and the 1991 Pinatubo explosions, the energy was obtained from the far-field records of acoustic-gravity waves (Donn and Balachandran, 1981; Tahira et al., 1996). An empirical equation proposed by Posey and Pierce (1971) was used to estimate the source energy from the microbarographic records:

$$E = \frac{13P[R\sin(l/R)]^{0.5}H(c\tau)^{1.5}}{4.22 \times 10^{15}} \quad (18.1)$$

where  $E$  is the energy release in the explosive equivalent of tri-nitro-toluene (TNT) in megatons (Mt),  $P$  is the first peak-to-peak amplitude (Pa),  $R$  is the

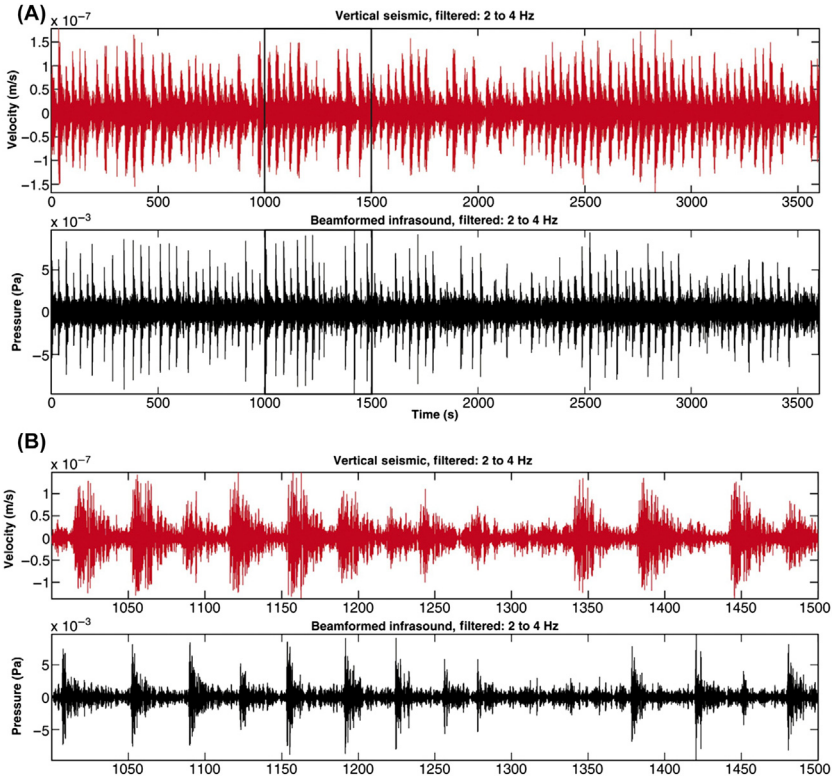


**FIGURE 18.23** Seismic and infrasonic recordings of “drumbeats” events observed on November 11, 2004, at Mount St. Helens at a distance of about 13 km from the summit. One hour of vertical seismic (top, *red*) and beam-formed infrasonic (bottom, *black*) recordings of “drumbeats” events are shown in (A). Both series are filtered 2–4 Hz. Box indicates the time range covered in (B). (B) displays 500 s of vertical seismic (top, *red*) and beam-formed infrasonic (middle, *black*) events from the time indicated in (A). From Matoza, R., Hedlin, M.A.H., Garcés, M.A. 2007. An infrasonic array study of Mount St. Helens. *Journal of Volcanology and Geothermal Research* 160, 249–262.

radius of the Earth (m),  $l$  is the great circle between the source and receiver (m),  $c$  is the sound speed (m/s), and  $\tau$  is the time interval (s) between the first and second peaks. The energy of the 1980 Mount St. Helens was estimated as 35 Mt of TNT and the energy of the 1991 Pinatubo explosions amounts to about 70 Mt of TNT.

Gorshkov and Bogoyavlenskaya (1965) calculated the energy  $E$  (in ergs) of air waves generated by the 1956 Bezymianny explosion from the near-field barograph records (45–780 km from the volcano) using the following equation:

$$E = 1.25 \times 10^{20} \sin \varphi \sum \frac{A^2}{2} \quad (18.2)$$

here  $\varphi$  is the distance from the explosion source in degrees and  $A$  is the amplitude of air wave in mbar. The energy of air waves was estimated as



$3 \times 10^{22}$  ergs (about 1 Mt of TNT). Gorshkov and Bogoyavlenskaya (1965) suppose that the energy of air wave represents about 0.1% of the total energy. Then the total energy of this explosion was estimated as  $1.2 \times 10^{24}$  ergs, or about 30 Mt of TNT.

>For comparison, the Little Boy atomic bomb dropped on Hiroshima on August 6, 1945, exploded with energy of about  $15 \times 10^{-3}$  Mt of TNT. The energy of the 1908 Tunguska meteoroid explosion was estimated as 30 Mt of TNT.

### 18.5.1.2 Small Explosions

The acoustic energy of small explosions was estimated for some explosive events considering that the acoustic waves were radiated spherically and that the total acoustic energy was proportional to the time-integrated squared excess pressure trace. Using the following equation:

$$E_{ac} = \frac{2\pi r^2}{\rho_a c} \int \Delta P^2 dt \quad (18.3)$$

where  $r$  is the distance between the source and receiver (m),  $\rho$  is the air density ( $\text{kg/m}^3$ ), and  $c$  is the sound speed (m/s), Johnson (2003) estimated the acoustic energy from the infrasonic signals recorded at the distances from 0.3 to 15 km for 12 small Strombolian explosive events recorded at 9 worldwide volcanoes. The acoustic energy, representing a very small portion of the total explosion energy (about 0.1%; Gorshkov and Bogoyavlenskaya, 1965), ranges between  $10^4$  and  $1 \times 10^8$  J. The estimations of the acoustic energy for some Strombolian and Vulcanian explosions at Villarica, Stromboli, Santiagito, and Fuego volcanoes span 5 orders of magnitude, varying from  $10^2$  to  $10^8$  J (Marchetti et al., 2009).

### 18.5.1.3 Pyroclastic Flows

Equation (18.3) may be used also for the estimation of the energy of infrasonic waves produced by moving pyroclastic flow. Yamasato (1997) obtained the energy of infrasonic waves from the 1993 Unzen pyroclastic flows of more than 3 km duration. It ranged between  $1.2 \times 10^6$  and  $4.7 \times 10^7$  J and was in almost the same order as the seismic energy.

## 18.5.2 Reconstruction of the Process of Dome Collapses and Pyroclastic Flow Movement

Figure 18.24 shows the distribution of infrasonic sources located during the dome collapses and following pyroclastic flows of Unzen volcano that occurred during June 1992–November 1993 (Yamasato, 1997). These sources of the infrasonic waves were estimated from waveform correlation and amplitude distribution. Pyroclasts began to excite infrasonic and seismic waves when they were fragmented by collision. Both seismic and infrasonic sources



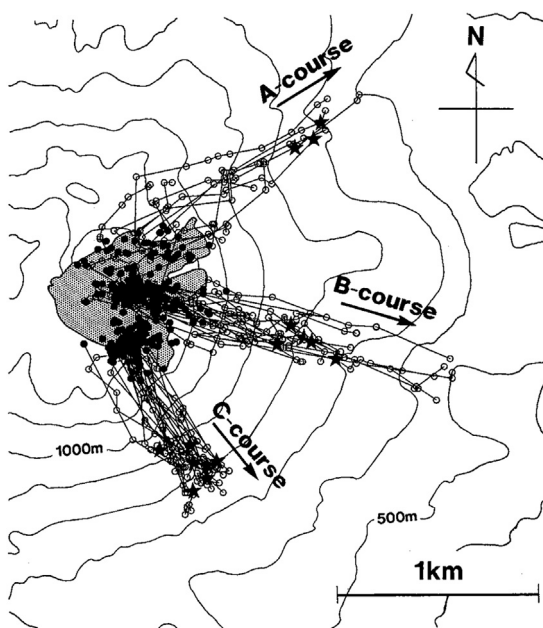


**FIGURE 18.24** Location of the sources of infrasonic signals from pyroclastic flows at Unzen volcano that occurred from June 1992 to November 1993. *Closed circles* indicate the sources due to dome collapses, and *open circles* indicate the sources of later phases that arise from the flow of pyroclastic. The hatched area indicates the locus of the lava dome. *Stars* indicate the locations where the amplitudes of the later phases of the seismic and infrasonic signals reached a peak. From Yamasato, H. 1997. *Quantitative analysis of pyroclastic flows using infrasonic and seismic data at Unzen volcano, Japan. Journal of Physics of the Earth* 45, 397–416.

were located near the front of pyroclastic flows. The infrasonic sources, corresponding to pyroclastic flows, show migrations along the pyroclastic flows pass trajectories. It allows determining the direction of a pyroclastic flow from analysis of infrasonic data even if visual observation is impossible.

### 18.5.3 Monitoring of Phreatic and Strombolian Explosions

The sequences of the infrasonic signals were recorded during the 2000 phreatic explosions at the Usu flank craters K and N (Yamasato et al., 2002). The pulses were excited at intervals of several seconds (Fig. 18.25A). Using the records from the infrasonic network, consisting of 10 stations and installed around the volcano (Fig. 18.25B), the source location of the pulses generated from two active craters K and N were identified. The amplitude of the infrasonic pulses reflected the activity of the phreatic eruption from these craters and they became smaller with the decline of the phreatic explosions at the active craters (Fig. 18.26).

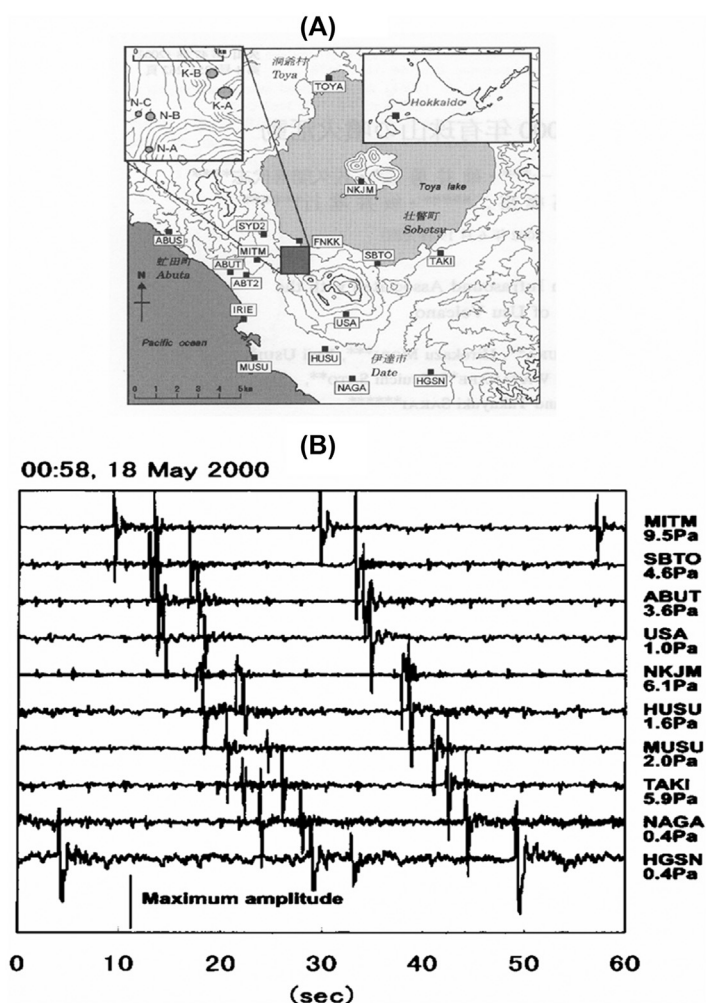


**FIGURE 18.25** Examples of infrasonic waveforms of the 2000 Usu phreatic explosions (B) recorded by the network (A) on May 18, 2000. Gray circles in (A) indicate the craters formed during the eruption. From Yamasato, H., Miyamura, J., Mori, H., Usui, Y., Sakuma, K., Watanabe, A., Sato, J., Takahashi, Y., Sakai, T. 2002. Eruptive activity inferred from infrasound associated with the 2000 eruption of Usu volcano. *Bulletin of the Volcanological Society of Japan* 47, 255–262.

Johnson et al. (2003) showed the utility of the infrasonic monitoring of Strombolian explosions at Karymsky, Sangay, Tungurahua, and other volcanoes. The infrasonic signals recorded at Tungurahua volcano indicated a continuous style of degassing, consisting of both explosive pulses and quasi-continuous “jetting.” At Karymsky volcano, where three types of acoustic signals were recognized (Fig. 18.27), the successive explosions, whether they

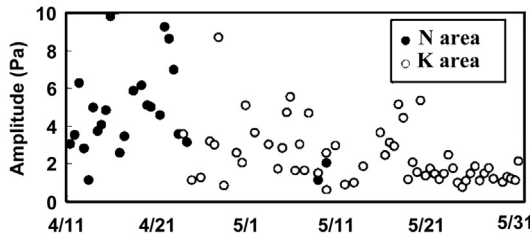


**FIGURE 18.26** Amplitude variations of the infrasonic pulses observed at ABUT station (see Fig. 17.22B) during the 2000 Usu phreatic explosions. *Closed and open circles indicate the infrasonic phases located at the N and K craters areas, respectively.* From Yamasato, H., Miyamura, J., Mori, H., Usui, Y., Sakuma, K., Watanabe, A., Sato, J., Takahashi, Y., Sakai, T. 2002. Eruptive activity inferred from infrasound associated with the 2000 eruption of Usu volcano. *Bulletin of the Volcanological Society of Japan* 47, 255–262.

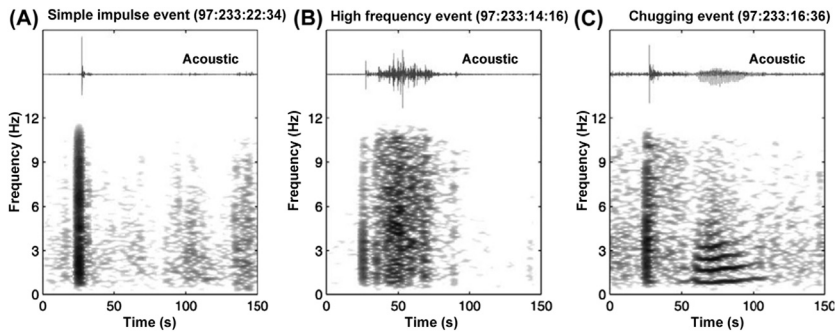


**FIGURE 18.27** Acoustic records of three types of Strombolian explosions at Karymsky volcano during 1997 activity and their spectrograms. (A) Simple impulse event, (B) high-frequency event, and (C) “chugging” event. Spectrograms are calculated with 10-s moving windows at 2-s increments and are bandpassed between 0.25 and 12.5 Hz. Code of date is yy:ddd:hh:mm, where ddd is a day from the beginning of the year. From Johnson, J.B., Aster, R.C., Ruiz, M.C., Malone, S.D., McChesney, P.J., Lees, J.M. Kyle, P.R. 2003. Interpretation and utility of infrasonic records from erupting volcanoes. *Journal of Volcanology and Geothermal Research* 121, 15–63.

were simple impulse, high-frequency, or chugging events, tended to vary according to the ease with which gas is able to escape to the free surface. The chugging at the Karymsky and Sangay volcanoes was associated with visible degassing.

Matoza et al. (2009b) investigated long-duration broadband infrasonic signals associated with sustained phreatic and magmatic Vulcanian, sub-Plinian, and Plinian explosions. Signals last many hours in duration in some cases (Fig. 18.28). Signals were attributed to a low-frequency form of jet noise. That is, the sustained jet flow in the gas-thrust region at the base of eruption columns is thought to radiate sound in a similar way to manmade

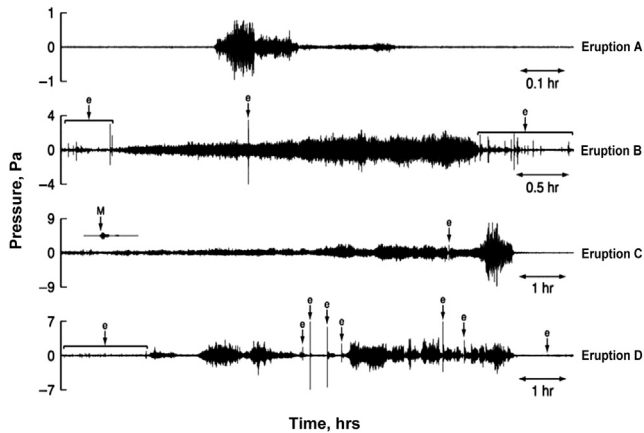


**FIGURE 18.28** Infrasonic signals recorded during Vulcanian and Plinian volcanic eruptions: (A) the March 8, 2005 Mount St. Helens eruption recorded at 13.4 km; (B–D) the Tungurahua eruptions recorded at 36.9 km (B, July 14, 2006–July 15, 2006; C, August 16, 2006–August 17, 2006; and D, February 6, 2008). Each trace represents a time domain beam across four array elements with unit gain. Discrete explosion occurrences marked by “e.” The signal labeled “M” on the third trace is the Mount St. Helens eruption (see A) plot at the same scale for comparison. From Matoza, R.S., Fee, D., Garcés, M.A., Seiner, J.M., Ramon, P.A., Hedlin, M.A.H. 2009b. Infrasonic jet noise from volcanic eruptions. *Geophysical Research Letters* 36, L08303.

jet flows, for example, those that come from jet engines or rockets. Since volcanic jet flows were very large in comparison to manmade jet flows, frequencies of radiation are much lower; hence Matoza et al. (2009b) considered them as infrasonic.

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# Seismic Monitoring of Volcanic Activity and Forecasting of Volcanic Eruptions

Chapters 5–10 of this book describe and discuss the properties of volcano-tectonic earthquakes and demonstrate the temporal and spatial regularities in the earthquake sequence development before volcanic eruptions. It allows forecasting of volcanic eruptions using seismic data. This chapter discusses the special aspects of seismic monitoring of volcanoes and the methodology of forecasting volcanic eruptions.

### 19.1 METHODOLOGY OF SEISMIC MONITORING OF VOLCANIC ACTIVITY

Seismic monitoring of volcanic activity serves to forecast an eruption and to issue an alert of a volcanic hazard to the public (Shimozuru, 1971; McNutt, 2000). Therefore the system of monitoring must be capable of precisely locating seismic events in the region of a volcano and able to quickly process the obtained information. To resolve this problem, a seismic network has to consist of a minimum of three to five seismic stations, with a spacing of a few kilometers and they need to be installed within 1–10 km from the active crater. Digital and analog seismic records are usually telemetered to a center of monitoring for processing.

#### 19.1.1 Seismic Networks Around Volcanoes

A seismic network around an active volcano, Volcán de Colima, Mexico, is shown in Fig. 19.1. It consists of four analog short-period vertical-component stations and five digital broadband three-component seismic stations distributed at a distance from 1.3 to 7.0 km and covering a broad azimuthal range. The stations are situated within the altitude range from 1700 to 4000 m; it is possible to locate earthquakes of magnitude  $-2$  to 4, which occur within the edifice of this volcano. The center of telemetrical monitoring is situated in

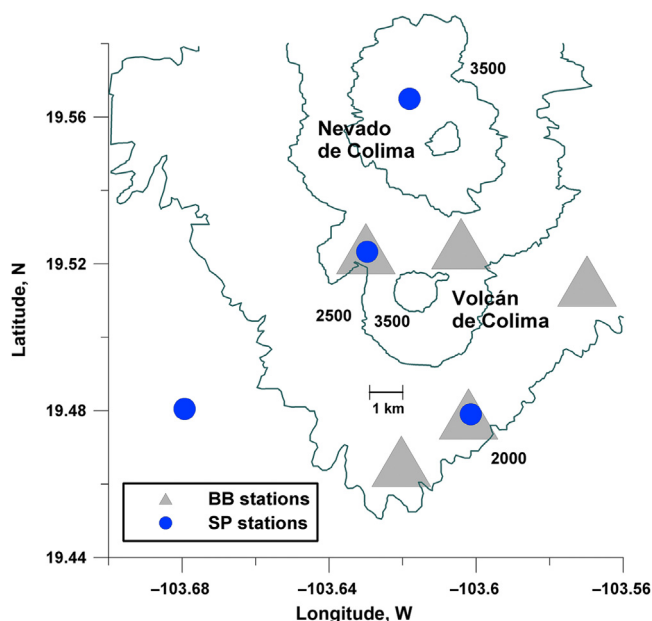


**FIGURE 19.1** The seismic network of short-period (SP) and broadband (BB) seismic stations installed around Volcán de Colima, Mexico. Shown are the contour lines for 2500, 3000, 3500, and 4000 m. *Courtesy of Colima Volcano Observatory.*

Colima city about 30 km from the stations. The analog data are transmitted by radio to the center where they are recorded on drums and stored as digital time series (Castellanos and Jiménez, 1995). This seismic network is of minimal capability; the destruction of any seismic stations during an eruption could stop the monitoring. The absence of three-component short-period stations lowers the precision of locations.

A good illustration of the importance of three-component stations and the use of both *P*- and *S*-waves for volcanic earthquake location is shown in Fig. 19.2. The 40 best-constrained earthquakes recorded at Etna volcano, Sicily are relocated in the absence of some essential data (Ferrucci, 1995). The seismic network consists of 13 stations, 11 of which are of three components. The location of events by all stations (Fig. 19.2B) constrains seismicity to within a focal volume of less than  $12 \text{ km}^3$  and confines it between sea level and a depth of 2 km. In Fig. 19.2C, the seismic cluster is shown after excluding the 2 three-component stations nearest to the crater, that were situated above the seismically active volume. It causes a dramatic spreading of the epicenters toward the southeast and an increase in the depth interval. In Fig. 19.2D, all stations are included but only *P*-wave data have been used; the result is only a weak effect on the epicentral distribution, but it strongly increases the depth interval.



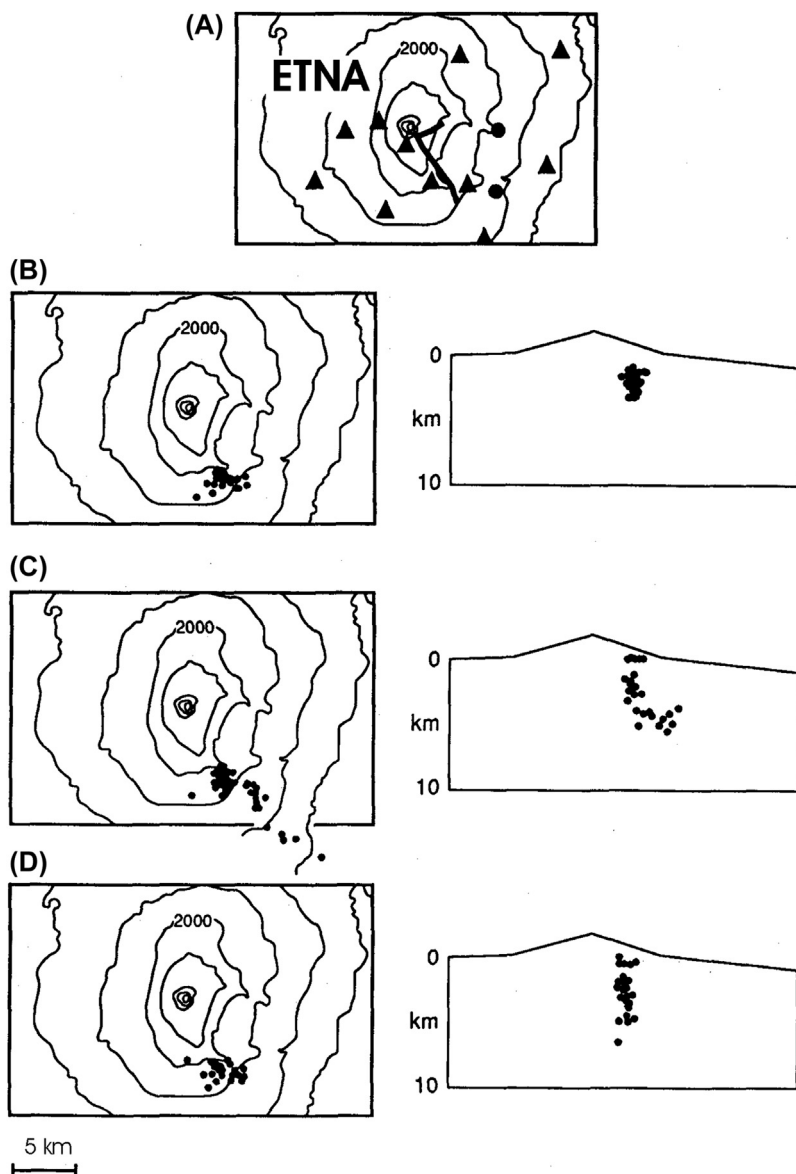


**FIGURE 19.2** Illustrations of the importance of the use of three-component stations in seismological routine and the adequate positioning of the stations according to the seismic volume at Etna volcano, Italy. (A) Seismic network (triangles, three-component stations; circles, one-component stations); (B) epicenters (left) and hypocenters (right) computed using all the *P*- and *S*-wave data; (C) epicenters (left) and hypocenters (right) computed without the two three-component stations nearest to the crater that was situated above the seismically active volume; and (D) epicenters (left) and hypocenters (right) computed using only the *P*-wave data. From Ferrucci, F., 1995. *Seismic monitoring at active volcanoes*. In: McGuire, B., et al., (Eds.), *Monitoring Active Volcanoes*. UCL Press, London, pp. 60–90.

### 19.1.2 Application of the Seismic Arrays for Study of Volcanic Seismicity

The seismic array is a system of linked seismographs arranged in a regular geometric pattern to increase sensitivity to earthquake detection. Originally built in the 1960s to detect and identify nuclear explosions, the seismic arrays became an important instrument in volcanic seismology. An example of the array application for the location of explosion earthquakes at Stromboli volcano was presented by Ibáñez *et al.* (2009). Two small-aperture, short-period seismic arrays were installed on the North and West flanks of Stromboli Island, at Labronzo and Ginostra sites, respectively (Fig. 19.3A). Ginostra array included 15 vertical components and 3 three-component Mark Products L4C seismometers, which have a natural frequency of 1 Hz. Labronzo array included 26 vertical components and 2 three-component Mark Products L15B seismometers with a natural frequency of 4.5 Hz.

Results indicated that the volume containing the source of the explosion quakes at Stromboli volcano was located beneath the crater area, and extended to depths not greater than 700 m below the surface (Fig. 19.3B). The horizontal extension, estimated at 80% probability, was between 0.5 and 1 km.

### 19.1.3 Initial Processing of Seismic Data

The processing of seismic data is realized in real time and afterward the earthquakes are localized. For real-time processing, digital and analog seismograms and spectrograms are used, as well as the compact presentation of seismic information with the real-time seismic amplitude measurement (RSAM; Endo and Murray, 1991) or real-time seismic energy measurement (RSEM; De la Cruz-Reyna and Reyes-Dávila, 2001) methodologies. Figs. 19.4, Fig. 19.5, and Fig. 19.6 show these three types of presentation of seismic information. The seismograms of volcanic earthquakes in an active period (Fig. 19.4A and C) give the general impression of the different types of seismic signals associated with volcano-tectonic earthquakes, explosions, rockfalls, or tremors. At the same time, it can be overfilled by a sequence of numerous and saturated signals. The spectrograms (Fig. 19.4B and D) help to separate low- and high-frequency events, to see the temporal variations of these signals.

The RSEM or RSAM curves give the characteristics of strain release associated with seismic activity. The RSAM or RSEM continuously samples the absolute amplitude (or square amplitude) of the seismic signal. The average of each signal is obtained by summing the measurements made in a certain time interval and dividing by the number of measurements (or unit time). These curves may serve as the characteristics of any stage in eruption development. It is illustrated in Fig. 19.5 where the great strain release is associated with the numerous seismic signals of rockfalls and pyroclastic flows during the extrusive stages of the lava dome growth (EXT1, EXT2,

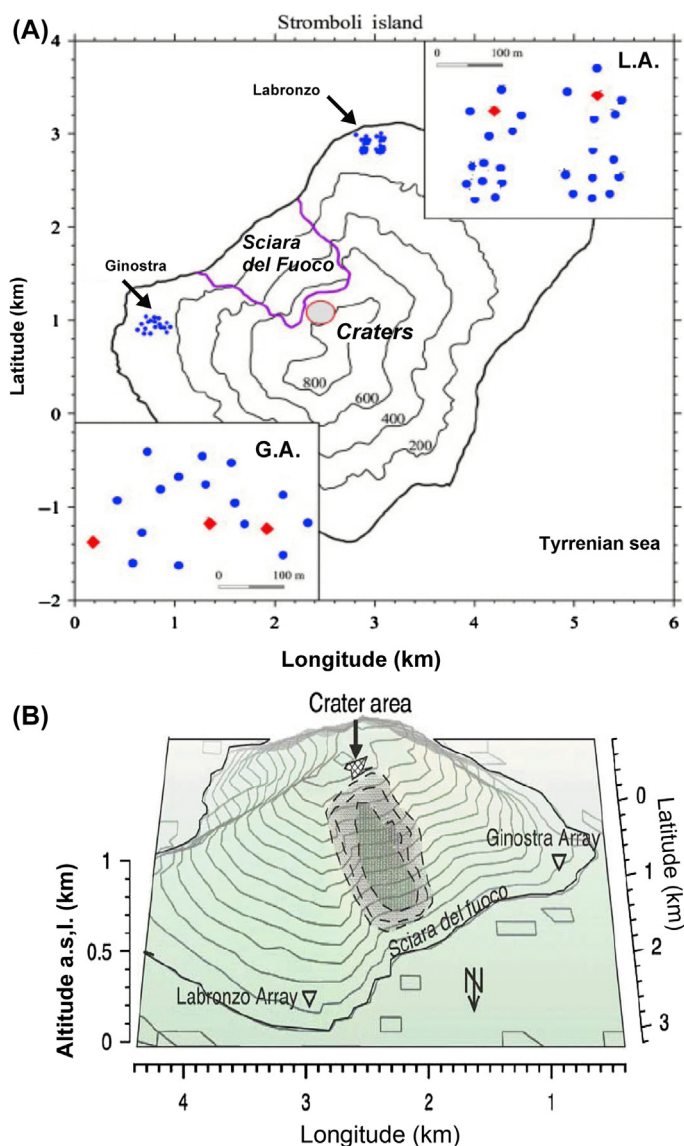


**FIGURE 19.3** (A) The position of the two seismic arrays (Labronzo, L.A., and Ginostra, G.A.) installed on Stromboli Island, Italy and the crater area. The two insets show the details of array configurations, in which full circles and diamonds indicate vertical component and three-component stations, respectively. The bold line bounds the sector of the graben Sciara del Fuoco. (B) Location of the source of explosion earthquakes obtained with two arrays. The array positions and the crater area are also indicated. From Ibáñez, J.M., Benítez, C., Gutiérrez, L.A., Cortés, G., García-Yeguas, A., Alguacil, G., 2009a. The classification of seismo-volcanic signals using Hidden Markov Models as applied to the Stromboli and Etna volcanoes. *Journal of Volcanology and Geothermal Research* 121, 15–63; Ibáñez, J.M., Saccorotti, G., La Rocca, M., Del Pezzo, E., Zuccarello, L. Alguacil G., 2009b. The use of seismic arrays to study the seismo-volcanic source. The example of Mt Etna and Stromboli Volcano. In: Bean, C.J., Braidon, A.K., Lokmer, I., Martini, F., O'Brien, G.S. (Eds.), *VOLUME Project VOLcanoes: Understanding Subsurface Mass Movement*. University College Dublin, Dublin, pp. 212–226.

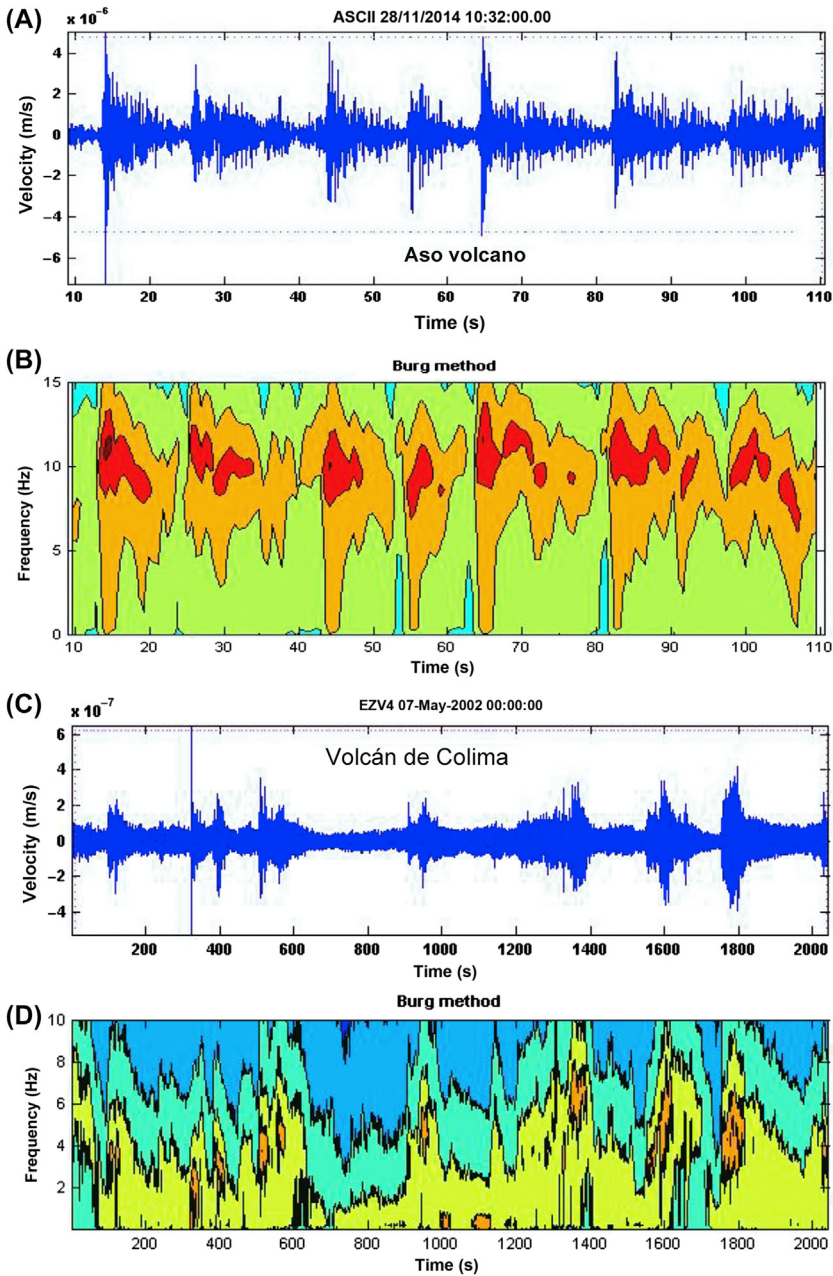


**FIGURE 19.4** Short-period seismograms recorded at the distance of 0.15 km from the crater of Aso volcano, Japan during the 2015 Strombolian activity (A) and recorded at the distance of 1.7 km from the crater of Volcán de Colima, Mexico during the 2002 lava eruption (C) and corresponding spectrograms (B, D). The spectrograms represent the time–frequency distribution with moving window of power spectral density calculated using Burg method. The numerous seismic signals were produced by explosions, rockfalls, and volcanic tremors. *Courtesy of Aso Volcano Museum and Colima Volcano Observatory.*

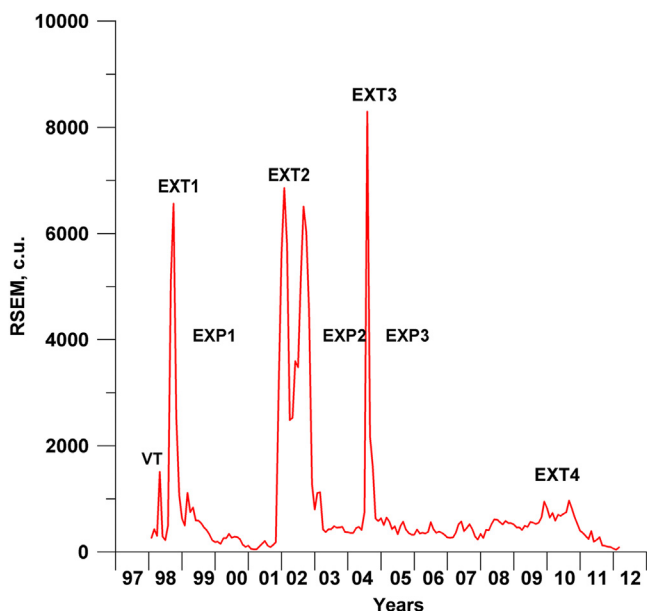


**FIGURE 19.5** Variations of RSEM during the 1997–2012 unrest at Volcán de Colima. The stages of volcano-tectonic (VT) earthquake swarms, lava extrusions (EXT), and Vulcanian explosions (EXP) are discriminated. *Courtesy of Colima Volcano Observatory.*

EXT3, and EXT4) at Volcán de Colima, while during the stages of explosive activity destroying these lava domes (EXP1, EXP2, and EXP3) the summary seismic amplitudes are low.

#### 19.1.4 Automatic Classification of the Seismic Signals

Automatic detection and classification problems play a critical role in the branch of seismology devoted to the continuous monitoring of seismogenic zones and active volcanic areas. The classification of seismic events should be done as close to real time as possible to reduce the response time in case of a volcanic alert. During the past decade, a new generation of automatic systems has been developed that allows very advanced analysis of the seismic records from the telemetered data in near real time (Esposito et al., 2006; Ibáñez et al., 2009).

The techniques developed to automatically discriminate between different kinds of seismic signals range from statistical analysis to cross-correlation and wavelet analysis. Limitations of deterministic models have led to more sophisticated methods that incorporate artificial neural networks and discrete hidden Markov modeling (HMM) tools to implement the classifier. These methods offer adaptation mechanisms to learn from the signals and assign model parameters (Ibáñez et al., 2009; Curilem et al., 2009).

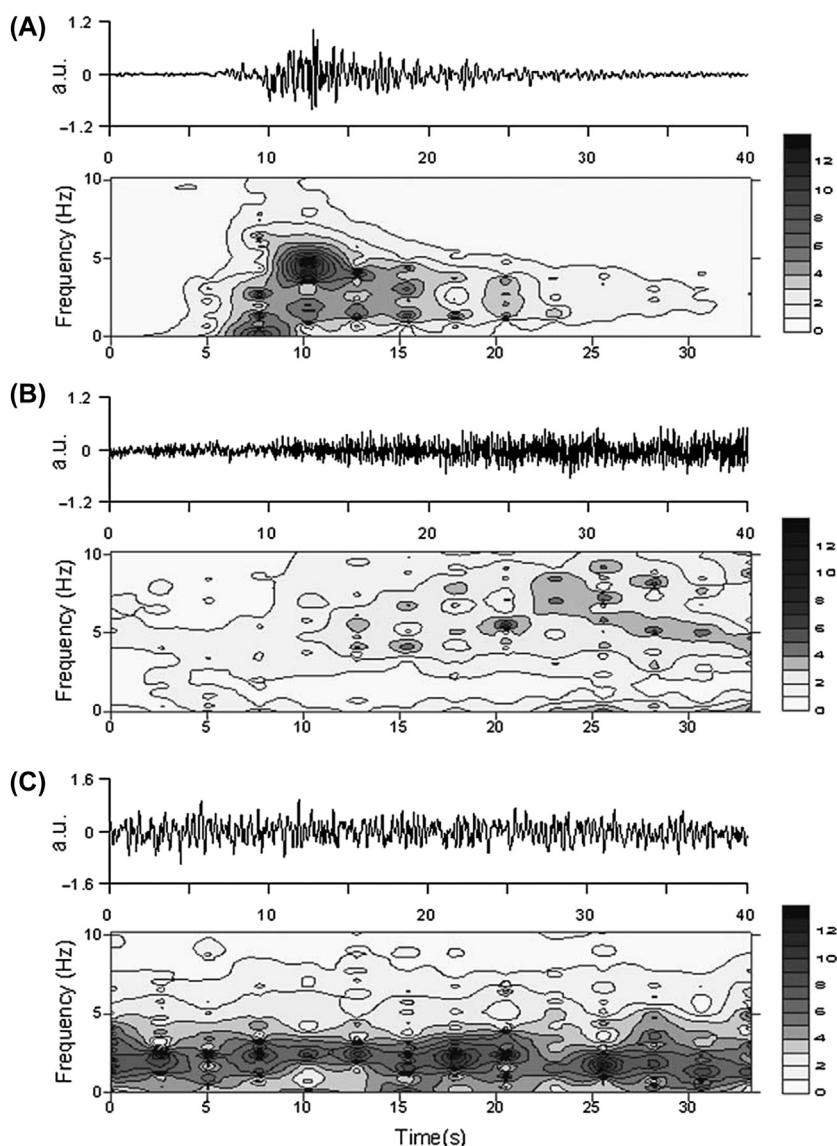


**FIGURE 19.6** Three typical signals used for automatic classification of the Stromboli seismic events are represented both in the time and time–frequency (spectrogram) domain. Whereas the explosion-quake (A) and the microtremor (C) signals exhibit a limited frequency content (mostly in the 1–6 Hz range), the landslide-related signal (B) has a broader spectrum. *From Esposito, A. M., Giudicepietro, F., Scarpetta, S., D’Auria, L., Marinaro, M., Martini, M., 2006. Automatic discrimination among landslide, explosion-quake, and microtremor seismic signals at Stromboli volcano using neural networks. Bulletin of the Seismological Society of America 96, 1230–1240.*

### 19.1.5 A Supervised Neural System

A supervised neural system was applied to discriminate among landslide, explosion-quake, and volcanic microtremor signals during the 2002 activity at Stromboli volcano (Esposito et al., 2006). The data were first preprocessed to obtain a compact representation of the seismic records. Both spectral features and amplitude-versus-time information have been extracted from the data to characterize the different types of events (Fig. 19.6). As a second step, they have set up a supervised classification system, trained using a subset of data (the training set) and tested on another data set (the test set) not used during the training stage. The number of hidden units and training cycles has been chosen empirically by trial and error.

The automatic system was able to correctly classify 99% of the events in the test set for both explosion-quake/landslide and explosion-quake/microtremor couples of classes, 96% for landslide/microtremor discrimination, and 97% for three-class discrimination (landslides/explosion-quakes/microtremor). A useful tool for analyzing classification results is the confusion matrix (Fig. 19.7), which contains information not only about the system performance but also about misclassified versus correctly classified data in the test set. For the explosion/landslide discrimination (Task A), all explosions were correctly classified, whereas just one landslide was misclassified (or confused) as an explosion. Similarly, for Task C (explosion/microtremor), two explosions were confused, whereas all microtremor signals were correctly identified. In Task B (landslide vs. microtremor), three landslides were confused and four microtremor events were misidentified. The network performed well for these two-class tasks, and it also did well in the three-class case, where explosions, microtremor, and landslides are jointly discriminated (Task D).

|                   |                  |             |  |                   |                  |             |     |
|-------------------|------------------|-------------|--|-------------------|------------------|-------------|-----|
| Task A            |                  |             |  | Task C            |                  |             |     |
| Network CM 99.61% |                  |             |  | Network CM 99.40% |                  |             |     |
| Explosion-quakes  | 156              | 0           |  | Explosion-quakes  | 167              | 2           |     |
| landslides        | 1                | 104         |  | microtremor       | 0                | 165         |     |
|                   | Explosion-quakes | Landslides  |  |                   | Explosion-quakes | Microtremor |     |
|                   |                  |             |  |                   |                  |             |     |
| Task B            |                  |             |  | Task D            |                  |             |     |
| Network CM 97.43% |                  |             |  | Network CM 98.39% |                  |             |     |
| Landslides        | 101              | 3           |  | Landslides        | 97               | 0           | 4   |
| microtremor       | 4                | 165         |  | explosion-quakes  | 0                | 167         | 1   |
|                   | Landslides       | Microtremor |  | microtremor       | 2                | 0           | 164 |
|                   |                  |             |  | Landslides        | Explosion-quakes | Microtremor |     |

**FIGURE 19.7** Confusion matrix on the test set for best network realization for each type of classification task at Stromboli. From Esposito, A.M., Giudicepietro, F., Scarpetta, S., D'Auria, L., Marinaro, M., Martini, M., 2006. Automatic discrimination among landslide, explosion-quake, and microtremor seismic signals at Stromboli volcano using neural networks. *Bulletin of the Seismological Society of America* 96, 1230–1240.



### 19.1.6 A HMM-Based Seismic-Event Recognition System

An HMM-based seismic-event recognition system decomposes the incoming seismic signal as a sequence of feature vectors. The length of the analysis window is such that the seismic waveform can be considered as stationary. The recognition system performs a mapping between a sequence of feature vectors and the corresponding sequence of seismic events. To obtain a useful data set, the continuous seismic waveform must be processed in two stages. The preprocessing stage is the selection of samples corresponding to the vertical velocity channel and conversion into a one-dimensional stream of 16-bit binary. The following feature extraction stage converts the volcano seismic signal into a parametric representation with less redundant information (Benítez et al., 2009).

Cortés et al. (2009) applied the automatic continuous HMM-based recognition system to the seismic events recorded at Volcán de Colima. A data set was collected at two short-period monitoring stations plus one broadband station mainly during the period of 2004–06, with some events from 1998. The acquisition system was attached to the widely used Earthworm software, developed by the US Geological Survey (USGS), which continuously digitizes the seismogram waveform. To set up the system and check database reliability, continuous close and blind isolated event recognition tests were performed: Close tests used the same data sets to train the models and to evaluate them, while in blind tests the training data set was not used for the evaluation. To improve the reliability of the results, and minimize the negative effect in the HMM from a lack of events, blind recognition is carried out by averaging the results of three tests where alternatively two data sets were used to train models and the third to evaluate. The recognition results are evaluated using the %correct and %accuracy rates (%Corr, %Acc) and the number of correctly recognized, substituted, and deleted events.

The following basic classes of seismic events were identified: volcanic tremor (T) of different types (T1–T3), volcano-tectonic (VT) earthquakes, regional earthquakes (REG), collapses (COL) or rockfalls, lahars (L), long-period events, and explosions (EXP). The confusion matrix, obtained for one of three data sets (Table 19.1) during continuous blind recognition of the shape of the seismic signals, was close to a diagonal matrix (which corresponds to a 100%Corr). The total number of recognized events in this set was 1575; the number of correctly recognized events was 1362; the number of substituted events was 84, and the number of deleted events was 129. The diagonal component of the matrix achieves a successful 81.65%Acc for Colima. Fig. 19.8 plots the continuous event classification carried out by the whole system during September 2008 to January 2009.

### 19.1.7 Location of Seismic Events

For a more detailed analysis of the situation, we need the maps and cross-sections of the focal distribution of the earthquakes. Fig. 19.9 illustrates these characteristics of the seismic regime for Volcán de Colima during the May



**TABLE 19.1** Confusion Matrix for Colima Events and Continuous Blind Test Results for Data Set 1

|     | SIL        | EXP        | LP         | REG        | T1        | T2       | T3         | COL        | LAH       | VT         | DEL | %COR   | %ACC  |
|-----|------------|------------|------------|------------|-----------|----------|------------|------------|-----------|------------|-----|--------|-------|
| SIL | <i>321</i> | 0          | 1          | 1          | 1         | 10       | 0          | 4          | 2         | 0          | 47  | 85.15  | 81.96 |
| EXP | 1          | <i>186</i> | 0          | 1          | 0         | 0        | 0          | 1          | 0         | 2          | 0   | 97.38  | 91.10 |
| LP  | 1          | 2          | <i>173</i> | 0          | 0         | 0        | 3          | 0          | 0         | 7          | 3   | 91.53  | 89.42 |
| REG | 0          | 3          | 0          | <i>111</i> | 0         | 0        | 1          | 1          | 0         | 0          | 0   | 95.69  | 89.66 |
| T1  | 0          | 2          | 0          | 10         | <i>50</i> | 0        | 0          | 0          | 0         | 1          | 13  | 65.79  | 57.89 |
| T2  | 1          | 1          | 0          | 2          | 0         | <i>6</i> | 0          | 0          | 0         | 0          | 0   | 60.00  | 60.00 |
| T3  | 2          | 1          | 5          | 1          | 0         | 0        | <i>194</i> | 1          | 0         | 1          | 29  | 82.91  | 79.91 |
| COL | 1          | 0          | 0          | 0          | 0         | 0        | 1          | <i>183</i> | 4         | 3          | 36  | 80.26  | 75.44 |
| LAH | 0          | 0          | 0          | 0          | 0         | 0        | 0          | 0          | <i>26</i> | 0          | 0   | 100.00 | 42.31 |
| VT  | 0          | 2          | 10         | 1          | 0         | 0        | 2          | 0          | 0         | <i>112</i> | 1   | 87.50  | 85.94 |

*Note:* The numbers of selected events of each of the types in diagonal components are indicated in *italics*.

*Source:* From Benítez, M.C., Lesage, P., Cortés, G., Segura, J.C., Ibáñez, J.M., de la Torre, A., 2009. Automatic recognition of volcanic-seismic events based on continuous Hidden Markov models. In: Bean, C.J., Braiden, A.K., Lokmer, I., Martini, F., O'Brien, G.S. (Eds.), The Volume Project. University College Dublin, Dublin, pp. 130–139.

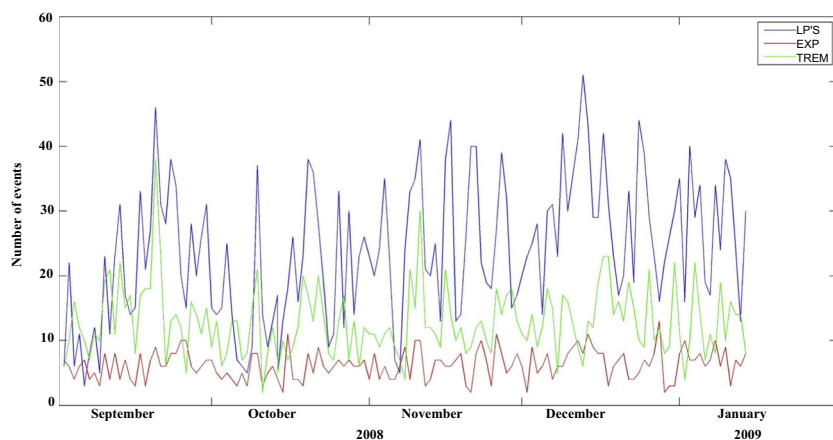


**FIGURE 19.8** Automatic continuous classification of events at Volcán de Colima between September 2, 2008, and January 14, 2009, showing long-period (blue line), explosions (red), and tremor (green) events activity. From Cortés, G., Arámbula, R., Álvarez, I., Benítez, C.M., Ibáñez, J.M., Lesage, P., et al., 2009. Analysis of Colima, Popocatepetl and Arenal volcanic seismicity using an automatic continuous hidden Markov model-based recognition system. In: Bean, C.J., Braiden, A.K., Lokmer, I., Martini, F., O'Brien G.S. (Eds.), *The Volume Project*. University College Dublin, Dublin, pp. 150–160.

2001 to June 2001 initial stage of lava dome growth. The events that occurred in May and June are marked by different symbols. This map and cross-section show that the June events became shallower and may reflect the intensification of lava dome growth.

## 19.2 APPLICATIONS OF VOLCANIC SEISMICITY TO THE FORECASTING OF VOLCANIC ERUPTIONS AND PREDICTING OF VOLCANIC HAZARDS

The case studies of Chapters 5–7 show that volcano-tectonic seismicity occurs before the majority of volcanic eruptions, indicating the site of the forthcoming event. F. Omori had written in May 1911, after the first historical instrumental study of seismic and volcanic activity during the 1910 Usu, Hokkaido eruption: “I believe the problem of forecasting of great volcanic eruption is, in some cases, not very difficult” (Omori, 1911–1913). He wrote later (Omori, 1914): “Unlike large destructive earthquakes, which originate along seismic zones, but are not repeated from one and the same center, the eruptions of a given volcano take place usually from the central crater or from new side vents, being anyhow thus localized to the mountain itself. Hence, it may be that the great outbursts from a given volcano at widely different epochs resemble each other more or less, both in the eruptive phenomena and in the precursory events.”

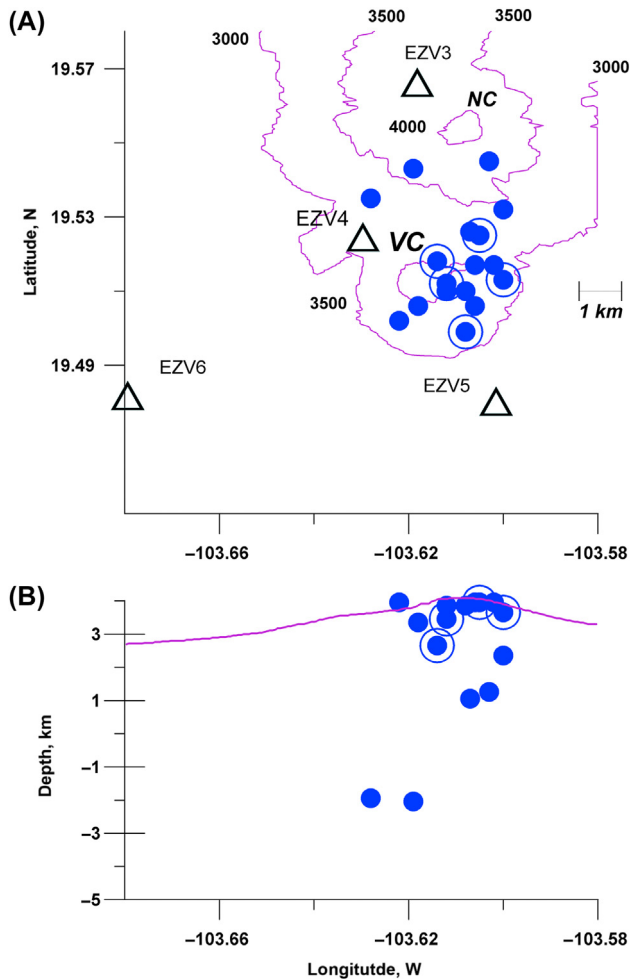


**FIGURE 19.9** Epicenters (A) and hypocenters (B) of small volcano-tectonic earthquakes located at Volcán de Colima, Mexico during May 2001 (filled circles) and June 2001 (double circles) at the initial stage of lava dome growth. The seismic stations are shown as *triangles*. The contour lines of 3000, 3500, and 4000 m are shown. NC, Nevado de Colima; VC, Volcán de Colima. *Courtesy of Colima Volcano Observatory.*

### 19.2.1 Methods Based on the Statistical Variations in the Parameters of Volcano-Tectonic Earthquakes

Minakami was a pioneer in the development of the seismic statistics method (Minakami, 1960, 1974). The studies were based on the monitoring of seismo-volcanic activity at the andesitic Asama volcano, Honshu. His famous diagram showing the relationship between the number of B-type volcano-tectonic

earthquakes and the daily frequency of volcanic events observed at this volcano (see Image 2.2 from Chapter 2) became classic. It was a typical example, in which premonitory earthquakes increased in number before an eruption. Minakami proposed the empirical laws (1960 and 1966) for predicting the Asama eruptions based on the 5-day frequencies of B-type earthquakes (Fig. 19.10). These laws predict that with an increasing number of B-type earthquakes, there is an increasing probability of an eruption.

The practical application of the 1960 formulas is shown in Fig. 19.11. The problem of forecasting was dealt with using the seismic frequencies of five successive days, as follows:

$$\begin{aligned} n_1 + n_2 + n_3 + n_4 + n_5 &= N_5 \\ n_2 + n_3 + n_4 + n_5 + n_6 &= N_6 \\ n_3 + n_4 + n_5 + n_6 + n_7 &= N_7 \end{aligned} \quad (19.1)$$

where  $n_1, n_2, n_3$ , and so on indicate seismic daily frequencies  $F_2$  of earthquakes and  $N_5, N_6, N_7$ , and so on indicate seismic frequencies  $N$  of five successive days. So, if the value of  $N$  on Day 1 is, for example, 800, the curve 1960 of Fig. 19.11 gives the probability  $P$  of an eruption within the following 5 days as 0.85, showing a very dangerous state of the volcano (Minakami, 1974).

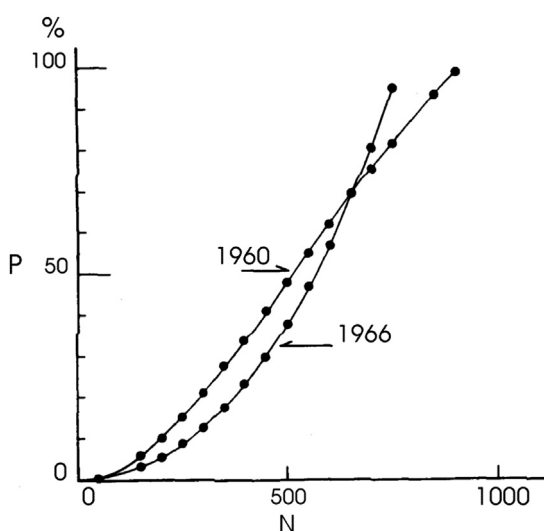


**FIGURE 19.10** The empirical curves formulas which were derived in 1960 and 1966 for predicting the Asama eruptions based on the 5-day frequencies of B-type earthquakes.  $P$  is the probability of eruptions in the next 5 days;  $N$  is the seismic frequency of the past 5 days. From Minakami, T., 1974. *Prediction of volcanic eruptions*. In: Civetta, L., Gasparini, P., Luongo, G., Rapolla, A. (Eds.), *Physical Volcanology*. Elsevier, Amsterdam, pp. 313–333.

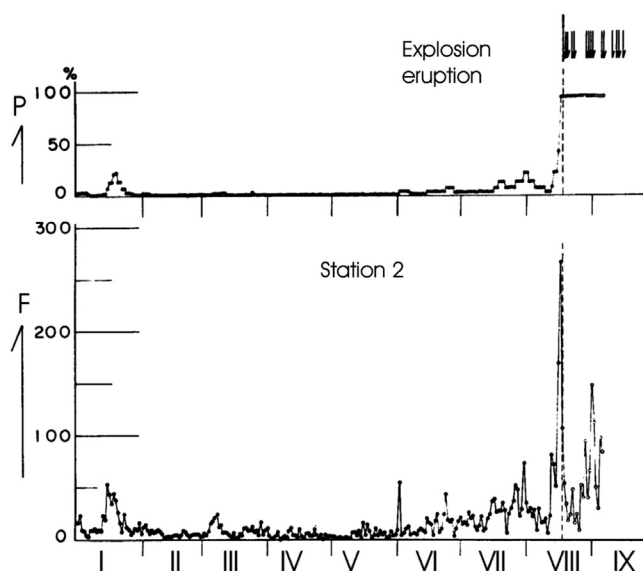


**FIGURE 19.11** Practical application of the 1960 formula to eruptions that took place in 1961, based on the daily frequency of *B*-type earthquakes. *Arrows* indicate the eruptions; *P* is the probability of eruption given from the 1960 formula; *F*2 is the daily frequency of *B*-type earthquakes originating from Asama. From Minakami, T., 1974. *Prediction of volcanic eruptions*. In: Civetta, L., Gasparini, P., Luongo, G., Rapolla, A. (Eds.), *Physical Volcanology*. Elsevier, Amsterdam, pp. 313–333.

Tokarev has developed this methodology for the forecasting of volcanic eruptions in Kamchatka using, not only the earthquake frequency but also the seismic strain release (Tokarev, 1963, 1971; 1978, 1983, 1985). For andesitic volcanoes, he proposed the hyperbolic law of cumulative strain release  $\varepsilon$ :

$$\varepsilon_t = \frac{at + b}{t - c} \quad (19.2)$$

where  $a$ ,  $b$ , and  $c$  are the coefficients for the given eruption and the time  $t$  is expressed in days. Eruptions occur in a period of time  $Tu \approx c$ . Fig. 19.12 illustrates this methodology. Coefficient  $c$ , which was large at the initial stage of the earthquake swarm, strongly diminished with the approach of the eruption. In practice, the date of the forthcoming eruption was estimated on Day 40 of activity, that is 1 week before the eruption.

This statistical methodology was strongly improved by including the material failure model (FFM) as a theoretical basis for the forecasting of an eruption (Voight, 1988; Voight and Cornelius, 1991; Cornelius and Voight, 1994, 1995). This FFM method is based on the empirical law describing the failure of materials. It is proposed that the acceleration (the second derivative) of any variable of state  $\Omega$ , which could be a possible precursor such as

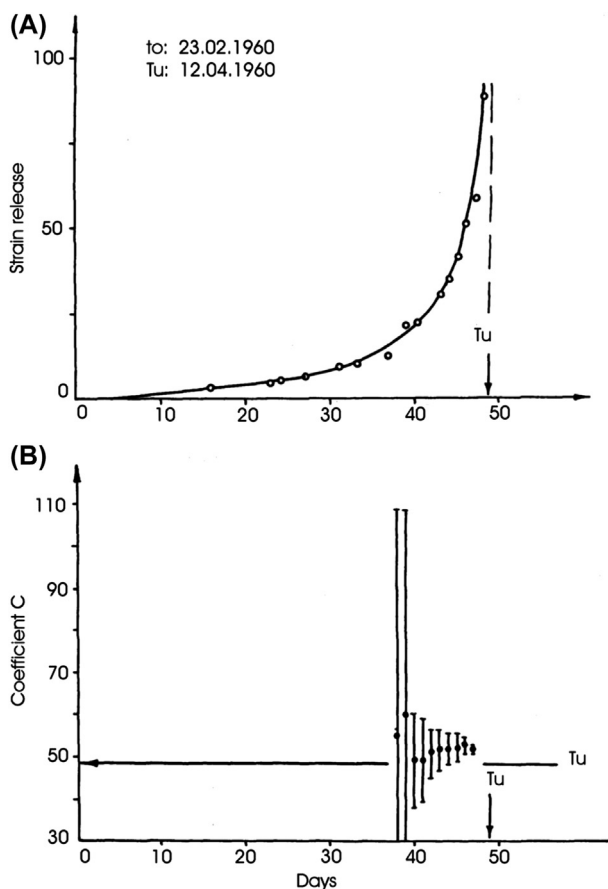


**FIGURE 19.12** Practical application of Tokarev's method for forecasting the eruptions of andesitic volcanoes. The variations of cumulative strain release (A) and coefficient  $c$  (B) before the April 12, 1960 eruption at Bezymianny volcano, Kamchatka are shown. Here  $t_0$  is the time of the first earthquake in the swarm. The strain release is expressed in units of  $[103] \times J^{1/2}$ ;  $c$  is expressed in days. Arrows show the time of the beginning of the eruption,  $T_u$ . From Tokarev, P.I., 1983. *Experience in predicting volcanic eruptions in the USSR*. In: Tazieff, H., Sabroux, J.-c., (Eds.), *Forecasting Volcanic Events*. Elsevier, Amsterdam, pp. 257–268.

seismicity, deformation, or gas emission, is proportional to the rate (the first derivative) of this variable to the order of  $\alpha$  (Eq. 19.3):

$$\Omega'' = A x \Omega'^{\alpha} \quad (19.3)$$

The empirical coefficient of proportionality  $A$  and the constant  $\alpha$  are dimensionless. For  $\alpha > 1$ , there is a time of singularity or the time at which the rate theoretically becomes infinite, and is, therefore, an upper bound to the time of failure (or the onset of the eruption, in our case).

The graphical technique of forecasting is based on a plot of the inverse rate of the variable of state versus time (Fig. 19.13). The cumulative values of RSAM or seismic spectral-amplitude measurement (SSAM) as well as cumulative coda duration or seismic amplitudes can be used as a seismic variable. When the cumulative seismic variable begins to increase sharply, the inverse rate of the variable goes to zero. The estimation of the time of the onset of an eruption is based upon successive estimations of the apparent failure times from the approximation of the inverse rate curve according to the following equation:

$$(\Omega''^*)^\beta = Ct + D \quad (19.4)$$

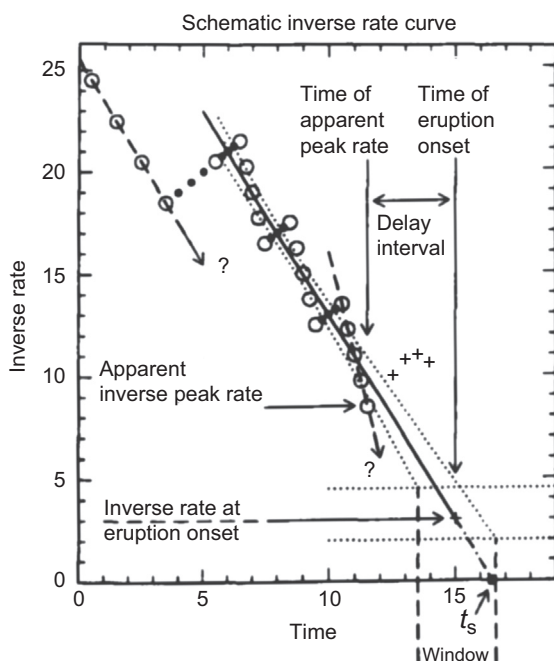


**FIGURE 19.13** Schematic inverse rate curve. Open circles define inverse downward (accelerating) rate trends, solid dots decelerating jogs. Decelerating jogs may cyclically alternate with accelerating jogs (dashes and arrows), which may be steeper than the overall trend (heavy solid line). Data points beyond the apparent peak rate were excluded from the analysis (crosses). Inverse rate trends can be fitted numerically and a data envelope (curved lines of small dots) can be constructed at a specified confidence level. The intersection of the data envelope with a range of expected rates at the time of eruption (horizontal lines of small dots), as derived from previous eruptions, may be used to define an eruption window (vertically dashed lines). Alternatively, the time of eruption may be defined in terms of an expected apparent peak rate, plus a delay interval. The time of singularity,  $t_s$ , is defined by the intersection of the curve fit with the abscissa, implying an infinite rate (square). From Cornelius, R.R., Voight, B., 1994. Seismological aspects of the 1989–1990 eruption at Redoubt volcano, Alaska: the material failure forecast method (FFM) with RSAM and SSAM data. *Journal of Volcanology and Geothermal Research* 62, 469–498.

where  $\beta = 1 - \alpha$ ,  $C$  is the slope of the inverse rate curve, and  $D$  is the intercept. The time of singularity, that is, the occurrence of an infinite rate, is found by extending the inverse rate data toward the time axis. A finite peak rate at the time of failure may be assumed, or estimated from experience from the earlier eruptions. This finite rate, plotted as a horizontal line in the inverse rate plot, can be used for the intersection with the extending trend of the inverse rate. The interval between the apparent time of failure and the time of eruption is termed the delay interval.

Fig. 19.14A–D shows the applications of the method for three types of seismic variables. These seismic data were observed before the October 22 (Julian day 295.5), 1986 eruptive event at the lava dome of Mount St. Helens. The inverse rates for all variables [coda duration (A), seismic wave amplitude (B), and RSAM (D)] were fitted to Eq. (19.4) using the linearized least-squares technique. The  $\alpha$ -values derived for the inverse rate curves are comparable (1.87 for coda accumulation, 1.67 for amplitude accumulation, and 1.38 for RSAM).

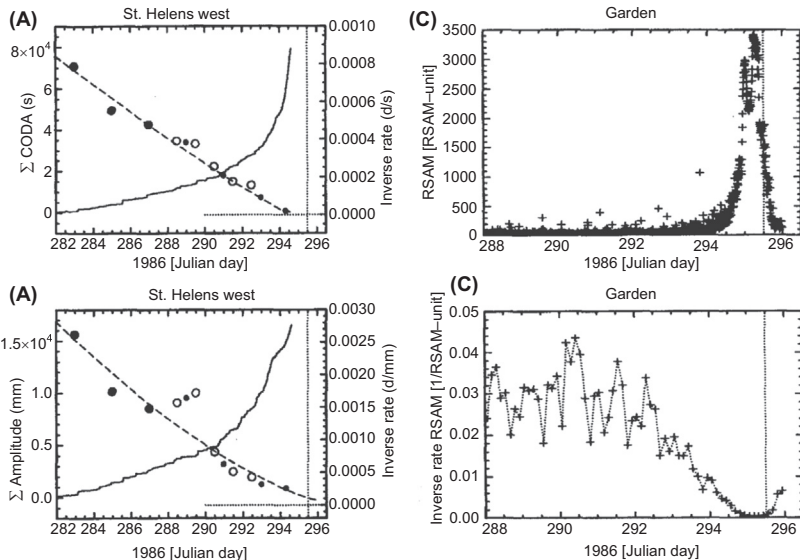


**FIGURE 19.14** Illustration of the FFM method applied to eruption prediction (October 22, 1986 Mount St. Helens volcano lava dome extrusion). Plots of cumulative coda duration (A), cumulative amplitude (B), and cumulative RSAM (C) are shown with their inverse rate curves (A, B, and D, respectively). Dashed lines in (A) and (B) show the fitted curves. ST HELENS WEST and GARDEN are seismic stations of the Cascade Volcano Observatory. Solid dots are the rates calculated over a period of 2 days; open circles are the mixed data set of rates calculated over periods of 2 and 1 days. Dotted vertical lines mark onset of extrusion. FFM, Material failure model; RSAM, real-time seismic amplitude measurement. From Cornelius, R.R., Voight, B., 1995. Graphical and PC-software analysis of volcano eruption precursors according to the material failure forecast method (FFM). *Journal of Volcanology and Geothermal Research* 64, 295–320.



The rates used for the curve fitting of A and B (solid dots in A and B) were calculated over time periods of 2 days, as daily accumulations. Calculating the rates over shorter time periods will usually increase data scatter but is feasible when the rate becomes larger. Examples (*circles*) in Fig. 19.14A and B are for accumulations per day calculated over periods of 1 day between October 14 and October 19 (Julian days 288–293). Distinct downward trends in the inverse plot are thus recognizable 2.5 days before the extrusion. The eruption began around Day 295.5. The forecasting windows were  $294.6 < 294.9 < 296.1$  for the A curve and  $295.0 < 297.2 < 301.6$  for the B curve. The forecasting window is based on a 97.5% confidence level for linearized fits according to FFM.

For the RSAM system (Fig. 19.14C and D), the data were arbitrarily averaged over 3-hour periods. The data window chosen for the RSAM analysis ends at the same time as one data window used for the analysis of coda and amplitude. The forecasting window for RSAM is  $296.3 < 296.8 < 297.6$ .

These methodologies work if the eruption occurs at the moment of maximum seismic strain release during volcano-tectonic earthquake swarm. The cases described in Chapters 5–7 show that this type of activity is common for andesitic and dacitic volcanoes but it is rare for basaltic volcanoes when the eruption may begin after the cessation of seismic activity.

The FFM methodology may be applied also to the development of a microearthquake swarm (Chapter 17.) before an explosion. Arámbula-Mendoza et al. (2011) applied this method to the microearthquake signals, preceding the September 16, 2005 explosion at Volcán de Colima, using seismic spectral energy measurement (SSEM) as a variable parameter. Fig. 19.15 illustrates the methodology. The time evolution of the inverse of SSEM that was computed for the spectral band between 1 and 3 Hz using a second-order Butterworth bandpass filter was studied using 10-minute-long windows. The explosion was preceded by a 3-day-long swarm of microearthquakes. During this period, the inverse SSEM displays a clear decreasing tendency. The adjusted straight line intersects the axis about 10 hours after the explosion (Fig. 19.15A).

### 19.2.2 Chronicle of Some Forecasting of Volcanic Eruptions Based on Seismic Monitoring

The methodologies described earlier are of interest but they are very unstable and may be applied to a limited number of well-known volcanoes. The problem is that the eruptive activity of a volcano changes the statistical regularities in its seismic activity without any announcement. As a result, the methodology, good serving during 3–5 years, may be wrong with the change in the eruptive behavior of the volcano. For example, the regularities in the seismic stress release before the 1957–70 eruptions at Bezymianny volcano, Kamchatka, associated with the lava dome growth episodes (Tokarev, 1981;

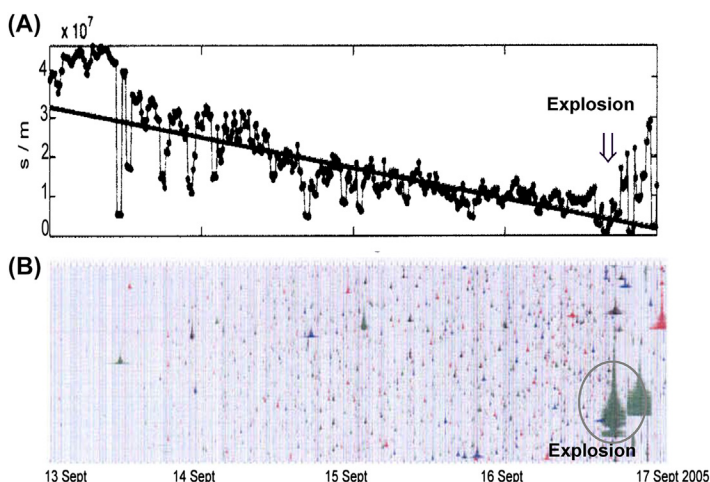


**FIGURE 19.15** Application of the material failure forecast method to Colima seismic data before the explosion of September 16, 2005. It shows the time evolution of the inverse of the seismic spectral energy measurement (SSEM) (A) and the seismic record of the sequence of microearthquakes (B). The *arrow* in (A) and *ellipse* in (B) indicate the actual occurrence of the explosion. From Arámbula-Mendoza, R., Lesage, P., Valdés-González C., Varley, N.R., Reyes-Dávila, G., Navarro, C., 2011. Seismic activity that accompanied the effusive and explosive eruptions during the 2004–2005 period at Volcán de Colima, México. *Journal of Volcanology and Geothermal Research* 205, 30–46.

Fig. 19.12 shows this regularity before the 1960 eruption) were changed at the beginning of 1970s. Chubarova (1995) wrote that the magnitudes of seismic events preceding the volcanic eruptions significantly decreased during 1971–94 (they were even completely absent before some of them), and the forecasting of eruptions at Bezymianny volcano became impossible. Therefore it is difficult to propose a general theory for volcanic eruption forecasting.

At the same time, Omori pointed out an advantage in predicting volcanic eruption: we know the place of the event. Therefore in short, the practical forecasting of volcanic eruptions is based on the simple fact that with a sharp increase of seismic activity near a volcano, we may have an eruption. The most effective forecasting of volcanic eruptions has had no special methodology; more important has been the human factor of experience. Of course, the forecasting will benefit from the use of additional types of monitoring such as deformation, chemical parameters, and so on. The author presented here the chronicles of successful monitoring and forecasting of volcanic eruptions.

### 19.2.2.1 The July 6, 1975 Eruption of New Tolbachik Volcanoes

This eruption in Kamchatka (see Section 5.3.1) was forecasted officially, with the publication of the date of eruption in the regional newspaper

(Tokarev, 1978). This forecasting was based on unusual seismic activity in an unusual place. The Tolbachik Valley, a region to the SW from Plosky Tolbachik volcano, where the swarm of volcano-tectonic earthquakes was localized, has been calm at least from 1960. Therefore the appearance of a seismic swarm could have been a signal of possible eruption. The swarm continued for 2 weeks giving time for the analysis of the situation and for correct decision. The forecasting of forthcoming eruptions was made by Tokarev according to the similarity in the seismic process near Tolbachik volcano with the seismic activity observed before flank eruptions of the Klyuchevskoy volcano.

The first seismic signals appeared on the records of seismic stations (the nearest of them was situated at a distance of about 50 km from the epicentral zone) on June 27 at 00:53, as a sequence of tremor-type vibrations (see Image 12.21 from Chapter 12: Volcanic Tremor). The first volcano-tectonic earthquake was recorded the same day at 01:09, and an intense swarm began at 09 hours (Fig. 19.16). During June 28 and June 29, hundreds of small earthquakes were recorded. The *P*- and *S*-wave arrivals of these events were sent from the seismic stations of the Kamchatkan network to the Institute of

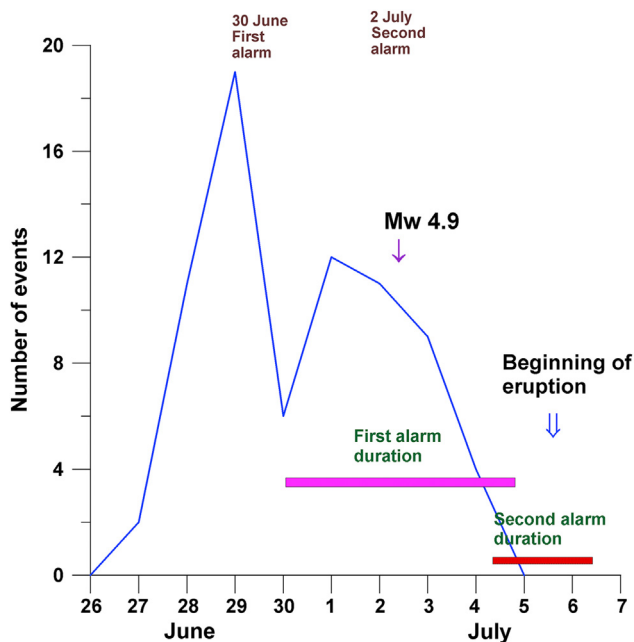


**FIGURE 19.16** A sequence of alarms before the 1975 Tolbachik eruption. *Period of alarms from Tokarev, P.I., 1978. Prediction and characteristics of the 1975 eruption of Tolbachik volcano, Kamchatka. Bulletin of Volcanology 41, 1–8. Data from Kamchatka Regional Earthquake Catalog.*

Volcanology, Petropavlovsk–Kamchatsky, every few hours by radio, and the seismic events were located in near real time. The epicentral area was estimated about 10 km SW from Plosky Tolbachik volcano. The development of seismic activity was similar to the activity preceding the lateral eruptions of the Klyuchevskoy volcano. On June 30, the administration of the Institute of Volcanology was notified that a new lateral eruption is expected SSW from Plosky Tolbachik during the period from June 30 to July 5 (First alarm).

The number of seismic events decreased on June 30; the forthcoming eruption was suggested to begin from July 2 to July 3. At the same time, two large ( $M$  4.9) earthquakes occurred on July 2, and a new pulse of seismic activity was recorded from July 2 to July 4. On July 3, Tokarev sent a message to the head of the Far Eastern Division of the USSR Academy of Sciences about the imminent eruption 10 km to SSW from Plosky Tolbachik volcano (Second alarm). A sharp decrease in the number of earthquakes on July 5 indicated that the eruption may begin during the nearest hours. It occurred on July 5 at 21:45 (GMT); local time July 6, 09:45 a.m.

The information for the citizens of Kamchatka about the forthcoming eruption was sent by Tokarev to the regional newspaper “Kamchatskaya Pravda” on July 1; the first short article was published on July 3. More detailed forecasting was published in the same newspaper on July 5.

The importance of this forecasting for the scientific study of the eruption was high. The first expedition of the Institute of Volcanology had arrived at the site of the forthcoming eruption on July 1; their camp was only 400–500 m from the place where a fissure opened on July 6. The eruption was observed by the volcanologists a few minutes after the opening of the fissure.

No people live within the Tolbachik Valley; no evacuation was made.

#### 19.2.2.2 *Forecasting of the June 15, 1991 Large Explosions of Mount Pinatubo, Philippines*

This forecasting was carried out also according mainly to a continuous increase in seismic activity beneath the volcano, after moderate manifestations of volcanic activity (Harlow et al., 1996). Detailed seismic activity is described in Section 7.1.4. During April–May, seismic activity was recorded mainly within a cluster located about 5 km NW of the volcano, and it was only 50% probable that a large eruption may occur. During the first week of June, the seismic activity beneath the volcano strongly increased, and the lava extrusion occurred on June 7. From June 8 to June 12, the majority of earthquakes occurred beneath the new dome; the tremor steadily increased from June 9. This seismic data together with the observations of an increase in  $\text{SO}_2$  flux from the volcano allowed the forecasting of the possibility of an increase in volcanic activity. This information was communicated to the officials, and the communities close to the volcano, as well as nonessential

personnel from Clark Air Base, were evacuated on June 10, 2 days before the first explosive eruption. The continuous increase in seismic activity beneath the volcano, including long-period and shallow volcano-tectonic earthquakes, and volcanic tremors showed that a climactic eruption was possible. During the morning of June 15, at 07:30 the monitoring team and the remaining military personnel abandoned Clark Air Base. The climactic eruption began at 13:42.

This forecasting of the volcanic eruption was realized only because the USGS volcanological team had great experience in the study of volcanic activity in other volcanic regions. They had no fixed methodology for this case, but they were prepared to try and understand the situation and make a decision.

### 19.2.2.3 *Forecasting of the November 20, 1998 Summit Eruption of Volcán de Colima, Mexico*

This forecasting (detailed seismic activity is described in Section 6.2.2) was based on the complex analysis of seismological and geochemical precursors (Reyes-Dávila and De la Cruz-Reyna, 2002; Zobin et al., 2002). The monitoring of Volcán de Colima was coordinated by the Scientific Advisory Committee (SAC), which was organized by Colima University from the very beginning of this volcanic unrest in October 1998 and later authorized by the Colima State Government. The SAC consisted of scientists from Colima University, UNAM, CENAPRED, and Guadalajara University, as well as state civil protection authorities, military officers, and foreign specialists, and served to advise the Colima State Government with regard to volcanic hazard mitigation. The SAC had meetings practically every week beginning in the spring of 1998; three of the meetings were held with the Governor of Colima State. The SAC was directed by the Head of Scientific Investigations of Colima University and by the Technical Secretary of the Civil Protection of Colima State. At each meeting, the SAC discussed the development of volcanic activity and prepared a report for public dissemination. The Bulletins of the SAC were published regularly on the Website of the Colima Volcano Observatory (CVO).

During 1998, two groups of SAC members saw indications in monitoring data for a future eruption of Volcán de Colima and expressed those concerns first in a long-term forecast, and later in a short-term prediction. The long-term forecast was based on observed increases of S/Cl and variations of deuterium D in gases from two summit fumaroles beginning in late 1997. Calling attention to those trends at the January 1998 session of the SAC, Y. Taran and J.C. Gavilanes forecast an effusive eruption “sometime in the nearest future, maybe this year.” On November 13, 1998, during the fifth seismic swarm (see Section 6.3.1), the SAC received a Memorandum prepared by two other members, S. De la Cruz-Reyna and G. Reyes-Dávila, in which they announced a high probability of a new eruption at Volcán de

Colima during the interval November 16 to November 18. Their short-term prediction was based mainly on seismic data and the FFM of [Cornelius and Voight \(1995\)](#). The predicted range of dates was estimated from three different models ([Fig. 19.17](#)). The authors of the prediction proposed three scenarios for the eruption, with the most probable scenario involving the formation of a new dome in the summit crater followed by lava flows and pyroclastic flows ([Reyes-Dávila and De la Cruz-Reyna, 2002](#)).

This Memorandum received the support of the SAC members, and on this basis recommendations for the State Government were elaborated. In response, on the morning of November 18, the Government of Colima State evacuated the  $\sim 180$  inhabitants of Yerbabuena, a village located 8 km SW from the crater of the volcano, and moved them to special prearranged shelters. On the same day, Jalisco State evacuated the  $\sim 120$  residents of Juan

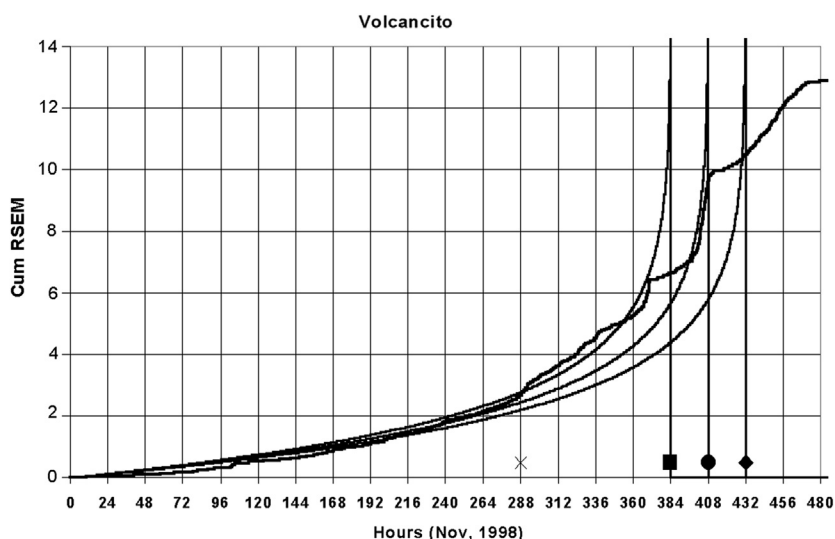


**FIGURE 19.17** The three material failure models used to support the forecast of an eruption at Volcán de Colima for the period November 16, 1998, to November 18, 1998. The thick irregular line represents cumulative 2-min averages of real-time seismic energy measurements (RSEMs), expressed as the square of the seismic signal amplitude  $\times 10^{-7}$  obtained during the first 20 days (480 h) of November 1998 from station Vulcancito, the nearest to the crater. The initial forecast, using the viscoelastic model, was issued late on November 12, 1998 (marked by “X”). Based on these 288 h of data, three sets of model parameters were used to define a forecast range for the failure time (times marked with a *square*, a *circle*, and a *diamond*). Around hour 371 (midday on November 16), the strain accumulation rate decreased for about 1 day and then resumed its increase. The regime of strain release changed again after hour 408 (late on November 17). A new lava dome was first seen growing in the summit crater at 07:30 h on November 20 (hour 463). *Modified from Reyes-Dávila, G.A., De la Cruz-Reyna, S., 2002. Experience in the short-term eruption forecasting at Volcán de Colima, Mexico, and public response to forecasts. Journal of Volcanology and Geothermal Research 117, 121–128.*

Barragan, located 10 km South-East of the crater. The inhabitants of Yerbabuena, Juan Barragan, and other proximal communities had been prepared for this action during many previous interactions with the Social Response Group of the CVO. These scientists had many discussions with the villagers during 1997–98 about the hazards associated with future eruptions of Volcán de Colima. As a result, the evacuation of November 18 to November 19 took place without complications. The lava eruption began during the night of November 19 to November 20.

#### 19.2.2.4 The March 31, 2000, Lateral Eruption of Usu Volcano, Hokkaido

This eruption (see Section 7.2.1) was the first case in Japan when a formal alert was issued by the Japan Meteorological Agency (JMA) before the eruption (Okada, 2002; JMA, 2003). Fig. 19.18 illustrates the process of forecasting. The number of volcano-tectonic earthquakes recorded at the volcano increased clearly during the morning of March 27. The Muroran Local Meteorological Observatory prepared the first report about this situation early on March 28. Two hours later, a Volcanic Advisory about the warning attention was issued.

Emergency meetings of the Coordinated Committee for Prediction of Volcanic Eruption of JMA were held on March 28 and March 29 to evaluate all data and information. Global positioning system (GPS) observations indicated that the volcano had inflated since March 29 (Takahashi et al., 2002). The Committee concluded that the eruption of the Usu volcano should be considered “imminent” within a few days. Based on this conclusion, a “volcanic alert” was issued on March 29 at 11:10 (Japan Standard time) by JMA. This alert signified that volcanic activity reached extremely high levels and urgent measures to prevent and mitigate danger to human lives are required. Based on the volcanic alert of March 29, local governments established a headquarters for volcanic disaster prevention and issued evacuation orders.

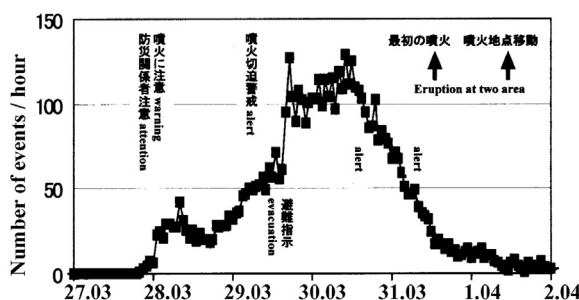


**FIGURE 19.18** Sequence of alarms before the 2000 Usu eruption. From Okada, H., 2002. Report on the 2000 Eruption of Usu and Investigation of Volcanic Risk. Institute of Seismology and Volcanology, Sapporo (in Japanese).



About 15,000 residents and tourists were evacuated from the area of volcanic danger by local governments without any confusion. After March 30, cracks and faults were also found in the northwestern foot and summit, indicating that crustal deformation had increased (JMA, 2003). The last alert was issued on March 31, a few hours before the eruption.

### 19.2.2.5 Forecasting of the October 1, 2004 Volcanic Eruption at Mount St. Helens, Cascades, and the Prediction of Associated Volcanic Hazards

This description of the volcano monitoring before an eruption is given according to the review article by Scott et al. (2008). It is of particular interest because it presents not only the analysis of variations in seismic activity but also shows the cooperation between the scientific, governmental, and media teams and shows the system of notification of ash hazards to aircraft.

A 2-day swarm of tiny (mostly  $M_d < 1$ ), shallow volcano-tectonic earthquakes at the volcano began early on September 23, 2004 (see Section 7.1.1; Fig. 19.19). The Information Statement by USGS–CVO and Pacific Northwest Seismic Network (PNSN) was issued at 18:00 Pacific Daylight Time (PDT) on September 23. The statement discussed the swarm’s similarity to the previous

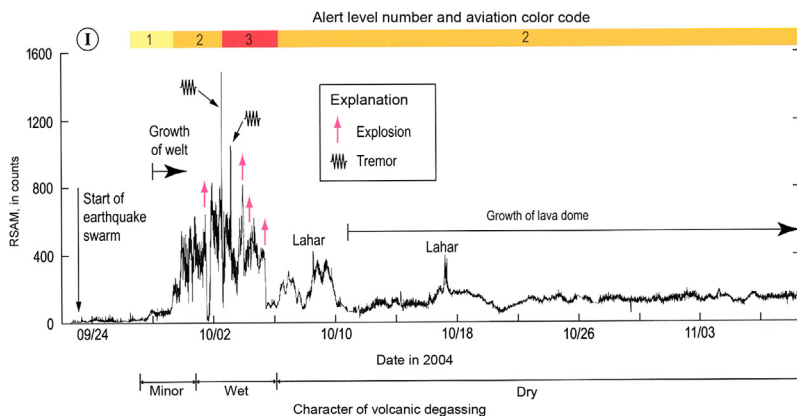


**FIGURE 19.19** Sequence of alarms before the 2004 Mount St. Helens eruption. Real-time seismic amplitude measurement (RSAM) during the first 6 weeks of unrest and eruption at Mount St. Helens served as a proxy for rate of seismicity. Times of explosions and periods of tremor are shown by symbols. *Modified from Moran, S.C., Malone, S.D., Qamar, A.I., Thelen, W.A., Wright, A.K., Caplan-Auerbach, J., 2008a. Seismicity associated with renewed dome building at Mount St. Helens, 2004–2006. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006. U.S. Prof. Paper 1750. Washington, pp. 27–60. From Scott, W.E., Sherrod, D.R., Gardner, C.A., 2008. Overview of the 2004 to 2006, and continuing, eruption of Mount St. Helens, Washington. In: Sherrod, D.R., Scott, W.E., Stauffer, P.H. (Eds.), A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006, U.S. Prof. Paper 1750. Washington, pp. 3–26.*



ones and surmised that this swarm might reflect heavier than normal precipitation in the preceding 4 weeks but considered eruption unlikely without significant further precursors. Beginning on the afternoon of September 25, earthquakes increased in magnitude ( $M_d \leq 2.8$ ), and by the following morning a total of 10 ( $M_d > 2.0$ ) earthquakes had been recorded—the most in a 24-hour period since the lava dome building ended in 1986. This increase in seismicity, along with the appearance of low-frequency and hybrid earthquakes, prompted a reassessment of the probability of hazardous activity. At 15:00 PDT on September 26, after notifying the Washington State Emergency Management Division and Giffort Pinchot National Forest, home of the Mount St. Helens National Volcanic Monument (GPNF), USGS—CVO and PNSN released a Notice of Volcanic Unrest (Alert level 1, Fig. 19.19) indicating that seismicity had surpassed background level and that the volcano was in a state that could evolve toward eruption.

The increase in alert level spurred several actions. The GPNF activated their Emergency Coordination Center and closed the south flank climbers' trail and other trails near the crater. Field crews took advantage of uncommonly clear autumn weather to begin the installation of additional seismometers and GPS receivers around the volcano. Aerial observations of the crater at this time showed no obvious changes. By September 29, seismicity had intensified to about 3 events/min with maximum magnitudes of  $M_d$  2.4–2.8, rising RSAM values, and hours-long periods during which repetitive earthquakes of similar waveform dominated the records. The well-known association of such seismicity with lava dome-building eruptions of the past few decades at Mount St. Helens and elsewhere raised eruption concerns further and prompted the issuance of a Volcano Advisory midmorning on September 29. The aviation color code was raised to orange (Alert level 2, Fig. 19.19) because of increasing concern that explosions could send ash to altitudes where air traffic would be affected. The Volcano Advisory signified that processes were underway that could lead to hazardous eruptive events, but not imminently.

The alert level rise to Advisory triggered the GPNF to activate a local Incident Management Team and to begin discussions with other agencies regarding a Joint Operation Center, which would be necessary if unrest continued to escalate or eruptions began. Intense media interest, directed chiefly toward USGS—CVO and the Monument, was beginning to adversely affect operations, so local and state emergency management agencies joined with USGS—CVO and the GPNF to begin organizing a Joint Information Center.

Real-time seismic amplitude values increased again on the evening of September 29, reflecting an increase in maximum earthquake magnitudes ( $M_d$  2.8–3.3) and rate of earthquakes of  $M_d > 2$  to about 1/min. The RSAM values fluctuated over the next 1.5 days. Flights during this time period again failed to detect significant amounts of volcanic gas. Data from a GPS receiver on the west side of the 1980s lava dome showed northward movement of several centimeters per day, consistent with a shove from a rising mass south of the dome.

The inaugural flight at Mount St. Helens of helicopter-mounted forward-looking infrared radiometer midday on October 1 was rewarded with close observation of the first explosion of the current eruption, which bored an ice-walled crater through the western part of the welt. An ash and vapor cloud rose about 2 km above the crater rim and drifted southwestward. Neither this nor other explosions over the next few days had any recognizable precursors. The rate of earthquakes rapidly increased again during the evening, and by late evening RSAM values exceeded those before the explosion.

For the next 2 days (October 2 to October 3), seismicity remained at a high level of several thousand events per day, punctuated by earthquakes of *Md* 3.5–3.9. Midday on October 2, a 50-min episode of energetic, relatively broadband seismic tremor, coupled with hints of volcanic gas during the previous few days and a continued high rate of growth of the welt, prompted USGS–CVO and PNSN to issue a Volcano Alert at 14:00 PDT. Because of concern over an imminent and potentially hazardous event, the GPNF evacuated the Johnston Ridge Observatory, 8 km north of the crater, the Washington State Department of Transportation closed State Route 504 at the Coldwater Ridge Visitor Center, and land and air space was closed within 8 km of the crater. The increased possibility of ash clouds reaching high altitudes where they could affect air traffic warranted raising the aviation color code to red (Alert level 3, Fig. 19.19).

Volcano monitoring did not give the exact date of the first explosion, but Alert level 2, pronounced on September 29, signified that the small explosive eruption was imminent. So its occurrence on October 1 was expected.

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# Seismic Activity at Dormant Volcanic Structures: A Problem of Failed Eruption

Numerous seismic events are located below a dormant volcanic structure where no seismic activity was observed for many years. What does it mean: a forthcoming eruption or only a seismotectonic episode? Volcano can be dormant for hundreds or thousands of years and then become active again; a new volcano may be born. We have a seismic sequence that frequently leads up to the eruption. But once in a while, we have a seismic sequence in a volcanic zone without any eruptive activity. The problem is: how to recognize if the seismic swarm announces the forthcoming eruption or if it is only a seismic crisis without volcanic manifestations, a failed eruption, or an instance in which magma has intruded to shallow depths (generally  $<2\text{--}3$  km below the surface), accompanied by anomalous seismicity, deformation, degassing, but ultimately fails to reach the surface, as was defined by [Moran et al. \(2011\)](#). In this chapter, the author gives some best-documented cases of seismic activity that were recorded at dormant volcanic structures and were not followed by an eruption; then some results of modeling of magma propagating to the surface and its arrest near the surface; and finally a short discussion of the problem of monitoring of the seismic activity at dormant volcanoes.

## 20.1 FAILED ERUPTIONS: CASE STORIES

The failed eruptions were recorded at different volcanic centers. The author discusses the occurrence of seismic swarms recorded without eruptions at large calderas, at stratovolcanoes, and in rift settings.

### 20.1.1 Failed Eruptions at Large Calderas

Large resurgent calderas are characterized by a slow posteruption upheaval of its floor. They form a broad topographical depression with a central elevated massif. Vertical uplift of more than 1 km often takes place ([Francis, 1998](#)). In this

section, two cases of seismic activity associated with the failed eruptions at large ancient calderas are described.

### 20.1.1.1 Long Valley Swarms

The large  $17 \text{ km} \times 32 \text{ km}$  Long Valley caldera (California, USA) was formed as a result of the voluminous Bishops Tuff eruption about 730,000 years ago. Resurgent doming in the central part of the caldera occurred shortly afterward, followed by rhyolitic eruptions from the caldera moat and the eruption of rhyodacite from the outer ring fracture vents, ending about 50,000 years ago. The caldera remains thermally active, with many hot springs and fumaroles (GVP, 2002).

Seismic swarms began in October 1978 and were accompanied by a broad uplift of the caldera floor and a local increase in fumarolic activity. The seismic activity culminated in a sequence of four magnitude  $M$  6 earthquakes that occurred beneath and to the south of the caldera from May 25, 1980, to May 27, 1980. Since a swarm in July 1984, most seismicity has been outside the caldera (Newhall and Dzurisin, 1988).

Most seismic events were typical tectonic earthquakes but also the episodes of spasmodic tremor and long-period events were observed (Cramer and McNutt, 1997). Fig. 20.1 shows the temporal characteristics of the 1989 swarm.

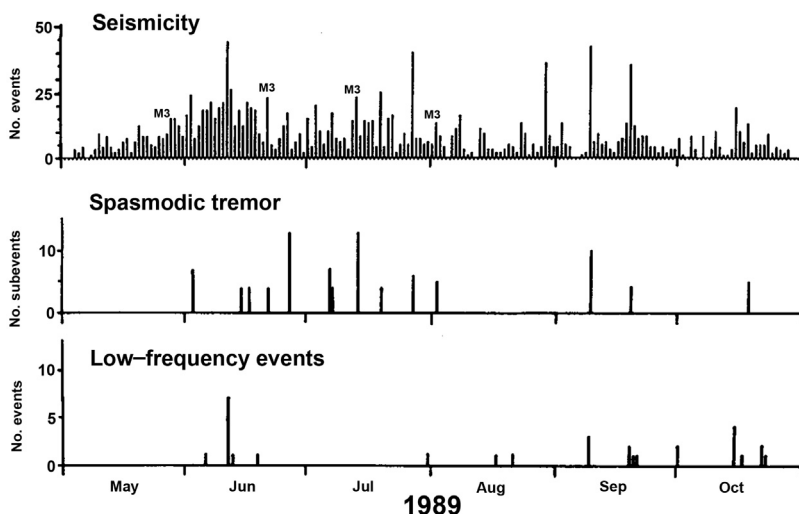


**FIGURE 20.1** Seismic events recorded during the 1989 swarm beneath Mammoth Mountain near Long Valley caldera, USA. The occurrences of magnitude ( $M$  3) events are shown. From Cramer, C.H., McNutt, S.R., 1997. Spectral analysis of earthquakes in the 1989 Mammoth Mountain swarm near Long Valley, California. *Bulletin of the Seismological Society of America* 87, 1454–1462.

This swarm occurred beneath Mammoth Mountain, which is situated at the SW rim of the caldera. It continued for about 6 months; more than 1500 events of magnitude 3 and less were recorded within the depth range of 2–11 km. Only 30 of them were identified as low-frequency events. Fourteen strong bursts of spasmodic tremor were also recorded. No regularities in the occurrence of low-frequency events and tremors were observed, and no volcanic eruption occurred during this period of caldera unrest.

### 20.1.1.2 Campi Flegrei Swarms

Campi Flegrei is a volcanic complex located near Naples, Italy. Eruption of about 80 km<sup>2</sup> of the Campanian Tuff about 35,000 years ago initiated the formation of the Phlegrean Fields, a complex of cones and craters within a 13 km diameter caldera (Francis, 1998). The last eruption was in 1538; it was preceded by an uplift that lasted over 30 years. After this eruption, the caldera has been dominantly sinking. A new period of uplift began near Pozzuoli in 1969. It continued for 4 years and reached 1.5 m. After a short gap, the uplift continued in 1982–84 with a maximum value of 1.8 m (Fig. 20.2). The uplift was accompanied by numerous small earthquakes (Barberi and Carapezza, 1996). Figs. 20.3 and 20.4 shows the temporal and spatial characteristics of the 1982–84 seismicity. During the period January 1983 to September 1984, more than 10,000 events were recorded with magnitude ranging between 0.1 and 4.2.

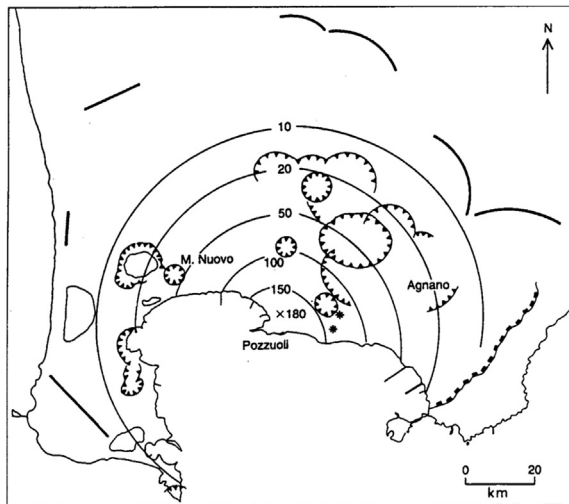


**FIGURE 20.2** Uplift of Campi Flegrei, Italy during the 1982–84 unrest. The curves are iso-lines of vertical uplift in centimeters. The cross near 180 cm indicates the point of maximum uplift. From Barberi, F., Carapezza, M.L., 1996. *The problem of volcanic unrest: the Campi Flegrei case history*. In: Scarpa, R., Tilling, R. (Eds.), *Monitoring and Mitigation of Volcano Hazards*. Springer-Verlag, Berlin New York, p. 771–786.

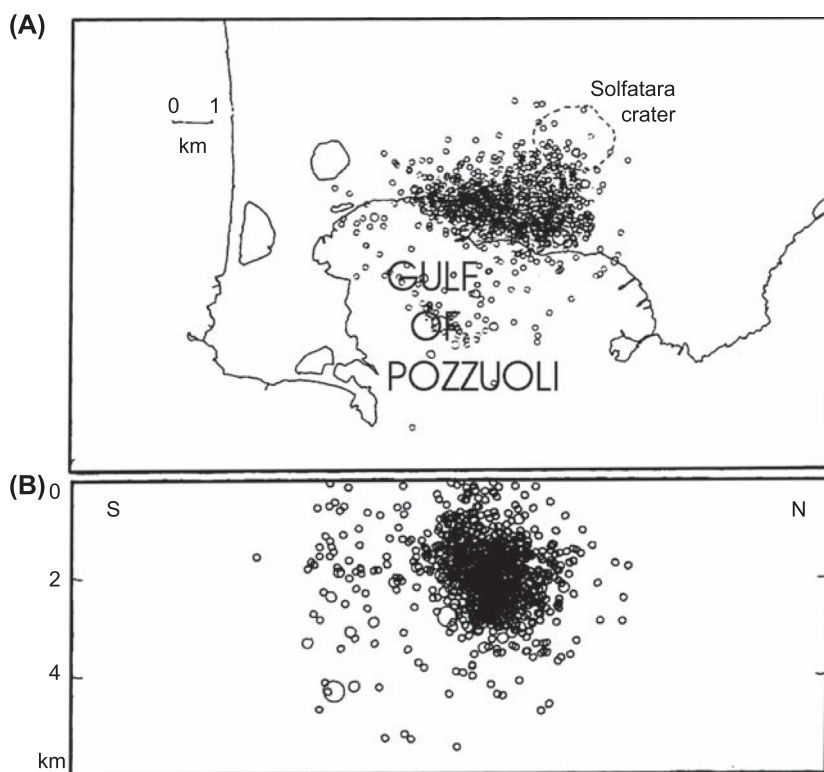


**FIGURE 20.3** The epicenters (A) and hypocenters (B) of earthquakes located during the 1982–84 unrest at Campi Flegrei, Italy. From Barberi, F., Carapezza, M.L., 1996. *The problem of volcanic unrest: the Campi Flegrei case history*. In: Scarpa, R., Tilling, R. (Eds.), *Monitoring and Mitigation of Volcano Hazards*. Springer-Verlag, Berlin New York, p. 771–786.

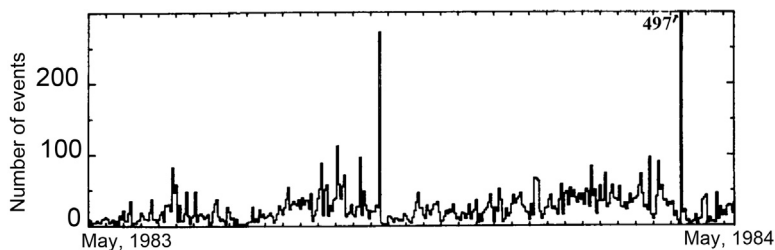


**FIGURE 20.4** Temporal variation in daily number of earthquakes recorded during the period May 1983 to May 1984 at Campi Flegrei, Italy. From Del Pezzo, E., De Natale, G., Martini M., Zollo, A., 1987. *Source parameters of microearthquakes at Phlegraean Fields (Southern Italy) volcanic area*. *Physics of the Earth and Planetary Interiors* 47, 25–42.



The earthquakes clustered within the depth range between 0 and 4 km; their epicenters were distributed within the uplift zone. Since January 1985, the area has undergone a phase of general subsidence interrupted by short-duration uplifts. The uplift phases were accompanied by seismic activity while the subsidence phases occurred aseismically (Del Pezzo et al., 1987). No volcanic eruption has occurred.

### 20.1.2 Failed Eruptions at Stratovolcanoes

In this section, three cases of seismic activity associated with failed eruptions at ancient stratovolcanoes are presented. The seismic activity was of different duration, from days and weeks to years.

#### 20.1.2.1 Matsushiro Swarms

The Matsushiro earthquake swarms (Central Honshu, Japan) began on August 3, 1965, with the earthquakes clustering within a 5-km radius zone around Mount Minakami that was not among the known active volcanoes of Japan. The swarm activity strongly increased in 1966 (Fig. 20.5) reaching about 7000 small earthquakes per day. From February 1966 to June 1966, strong deformations accompanied by ground fissures and spring water were

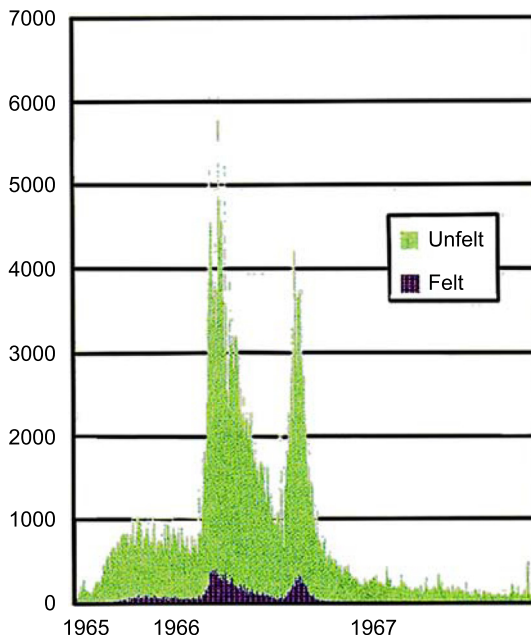


**FIGURE 20.5** Daily frequency of earthquakes during the Matsushiro earthquake swarm. From *Seismic Activity in Japan, 1998*. Earthquake Research Committee, Tokyo, 222 p.

observed. From July 1966 to December 1966, the spring water and the expansion of ground fissures caused numerous landslides.

The area of the epicentral zone spread significantly during the first 2 years of activity (Fig. 20.6). After June 1967, activity quickly dissipated and completely stopped by the end of 1970. More than 700,000 events were recorded during this swarm and more than 60,000 of them were felt. The largest of these earthquakes was with magnitude  $M$  5.4; the total energy of the swarm events was equivalent to the magnitude  $M$  6.4 earthquake. No eruption was observed (Seismic Activity in Japan, 1998).

### 20.1.2.2 Aag Swarm

The swarm of 113 earthquakes, occurring within the edifice of dormant, 2319-m high, andesite-basaltic stratovolcano Aag, Southern Kamchatka, began on August 10, 1966, and continued for 65 hours (Gorelchik, 1969). Aag belongs to the Avachinsky-Koryaksky group of Middle Pleistocene–Holocene volcanoes (Masurenkov et al., 1991).

A total of 46 events were located during the swarm (Fig. 20.7A). Epicenters clustered within the compact area of about 80 km<sup>2</sup>. They occurred at a depth between 5 and 10 km. A maximum event with magnitude  $mb$  3.5 was recorded during the peak in the number of seismic events, 5 hours after

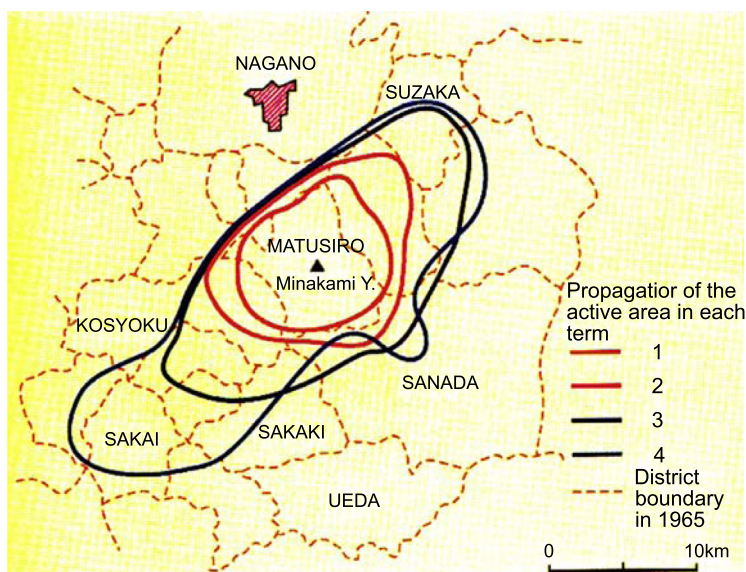


**FIGURE 20.6** Spreading of the source region during the Matsushiro swarm up to (1) February 1966, (2) July 1966, (3) December 1966, and (4) May 1997. From *Seismic Activity in Japan*, 1998. Earthquake Research Committee, Tokyo, 222 p.

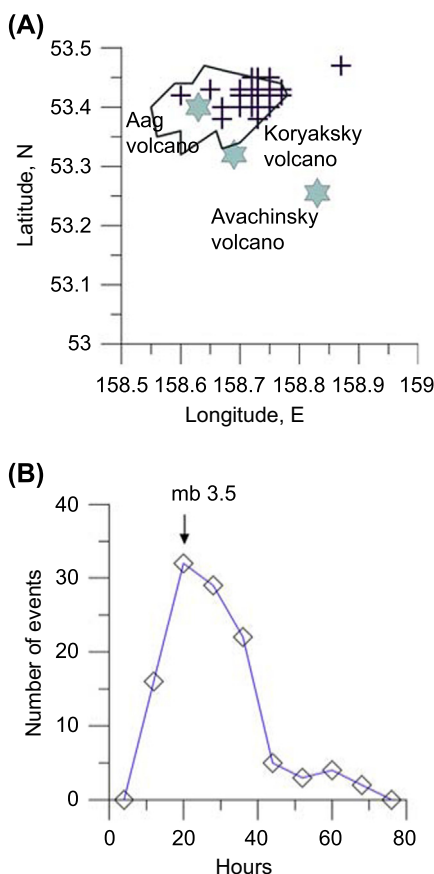


**FIGURE 20.7** Earthquake epicenters (A) and variations in the number of events (B) during the 1996 Aag swarm, Kamchatka. Stars show the volcanic centers; crosses mark the epicenters. Contours of the volcanic edifice of Aag volcano are shown by the solid line. Data from the Kamchatka Regional Earthquake catalog.

the beginning of the swarm (Fig. 20.7B). Seismic records displayed the typical waveforms of volcano-tectonic earthquakes with clear arrivals of *P*- and *S*-waves (Fig. 20.8). No eruption occurred. No new swarm was observed during the last 50 years.

### 20.1.2.3 Asacha Swarm

Intense swarm of volcano-tectonic earthquakes occurring within the edifice of the andesitic, 1910-m high, complex volcano Asacha, Southern Kamchatka, was recorded between March 7, 1983 and March 30, 1983 (Zobin et al., 1986). Asacha represents the complex massif of four Pleistocene-to-Holocene strato- and shield volcanoes located within a distinctly fault-bounded crustal

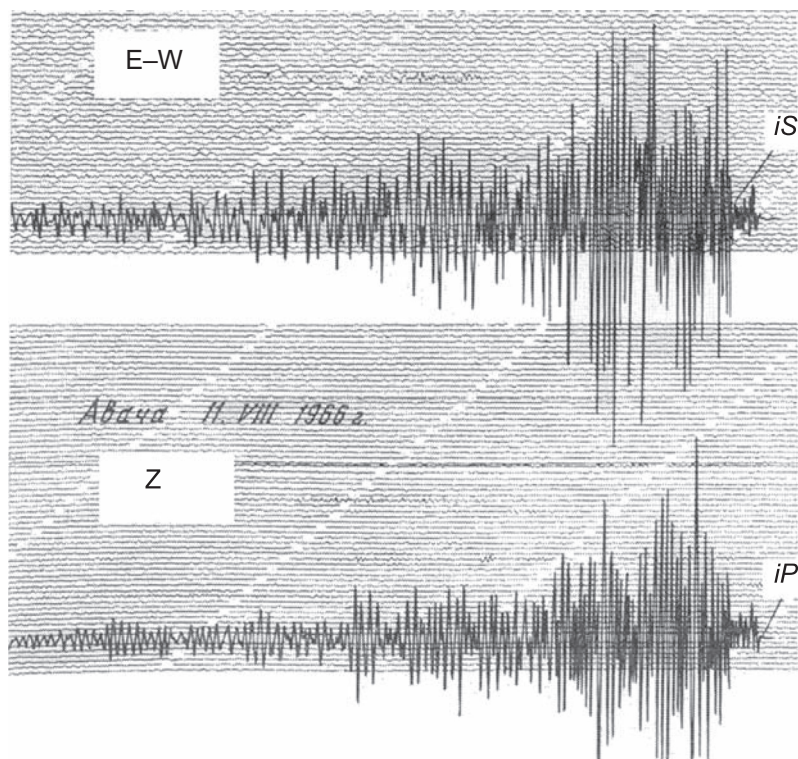


**FIGURE 20.8** Seismic record (displacement, horizontal, E–W, and vertical, Z, components) of one of the largest earthquakes of the Aag swarm recorded by the short-period station at a distance of 17 km. *Courtesy of the Institute of Volcanology.*

block. Ten lava domes dot the flanks of the Asacha volcanoes. Most of them were formed during the Pleistocene, but some may be early Holocene in age. During the Holocene, basaltic cinder cones and lava flows related to regional volcanism erupted along the western and southern flanks of the Asacha complex (GVP, 2011).

A total of 390 events were located during the swarm (Fig. 20.9A). They filled the area extending to the east from the main volcano summit zone and covering the central part of the volcanic complex edifice. The majority of events (77%) occurred within the depth interval between 5 and 20 km. Three peaks in the swarm activity were observed (Fig. 20.9B). During the peak activity of March 13, five earthquakes with magnitudes  $m_b$  (US Geological Survey—USGS) between 4.5 and 5.0 were recorded. These large events were located along the western border of the epicentral area including the main summit zone of the volcano. No eruption occurred. No new swarm was observed during the following 30 years.

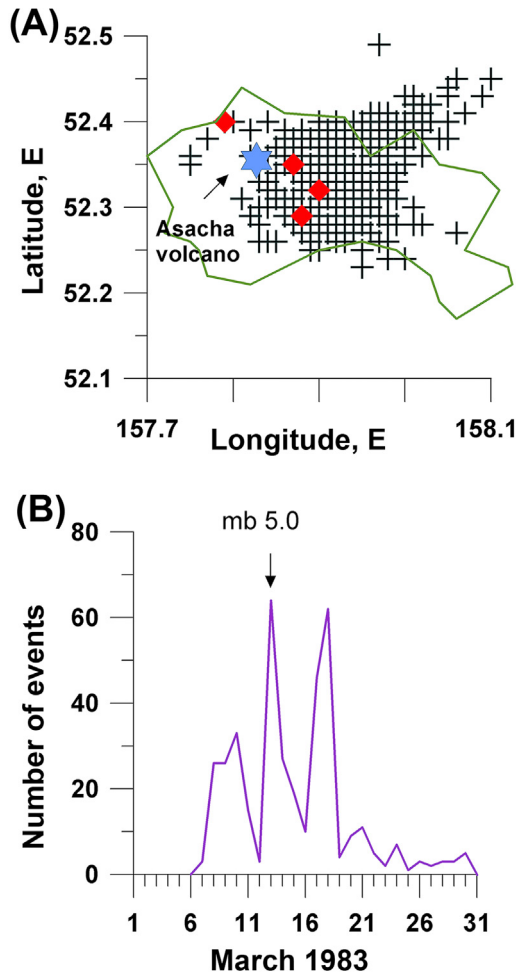


**FIGURE 20.9** Earthquake epicenters (A) and variations in the number of events (B) during the 1983 Asacha swarm, Kamchatka. The epicenters of largest earthquakes with magnitude *mb* 4.5–5.0 are shown with filled diamonds. Star shows the volcanic center; crosses mark the epicenters. Contours of the volcanic edifice of Asacha volcano are shown by the solid line. Data from the Kamchatka Regional Earthquake catalog and US Geological Survey catalog.

### 20.1.3 Failed Eruptions in Rift Settings

In this section, the seismic activity associated with two failed eruptions in a rift setting is described. One of the cases was observed in the northern volcanic rift zone in Iceland, and the other, in the Arabian rift zone.



### 20.1.3.1 Icelandic Rift Zone Swarm

Since February 2007, over 10,000 microearthquakes, ranging in size from  $ML$  0.9 to 2.2, have been detected near Mount Upptyppingar (Fig. 20.10).

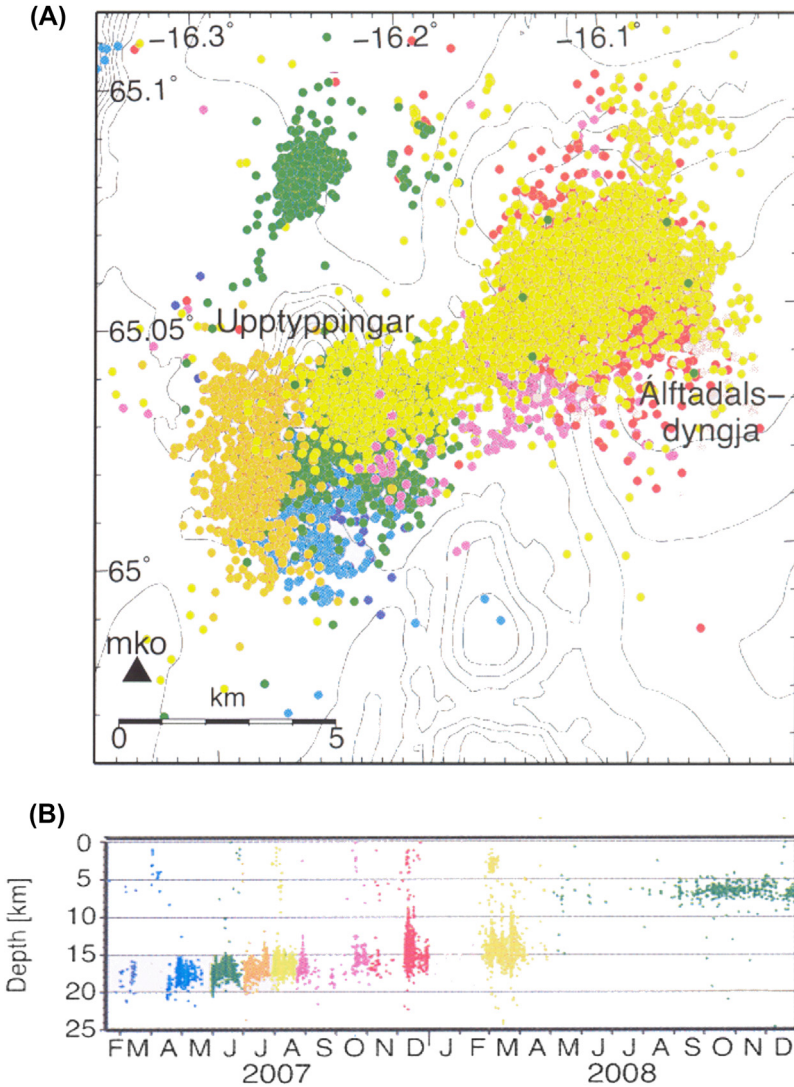


**FIGURE 20.10** Spatial and temporal view of seismicity at Upptyppingar and Álfadalsdyngja, Iceland: (A) earthquake epicenters and (B) temporal variation of hypocenters. Different colors correspond to temporal development of the swarm. The nearest seismic station MKO is shown with triangle. From Geirsson, H., Jakobsdóttir, S.S., Guðmundsson, G.B., Roberts, M.J., Deciem, J., 2009. A deep-seated magmatic intrusion at Upptyppingar, Iceland, during 2007 and 2008. In: Bean, C.J., Braiden, A.K., Lokmer, I., Martini, F., Ó'Brien, G.S. (Eds.), *The Volume Project*. University College Dublin, Dublin, pp. 2–12, with modifications.

Upptyppingar is located within the northwestern edge of the fissure swarm that occupies part of the Kverkfjöll volcanic system, within the Northern Volcanic Zone of Iceland. Rising to  $\sim 400$  m above the surrounding landscape, the mountain comprises hyaloclastic deposits formed in a subglacial eruption toward the end of the last glaciation, over 10,000 years BP.

The seismic records were characterized by impulsive *P*- and *S*-wave onsets. Over an 11-month period, the seismically active region migrated from Upptyppingar, spreading  $\sim 12$  km east–northeast to the Álftadalsdyngja lava shield. Starting from a depth of 18 km below sea level (b.s.l.), the seismicity migrated in both up- and down-dip directions and also laterally before stopping at 13 km depth. In addition to sustained earthquake activity, significant horizontal displacement was registered at two permanent global positioning system stations in the Upptyppingar region (Jakobsdóttir et al., 2008; Geirsson et al., 2009; White et al., 2011).

The seismicity was supposed (White et al., 2011) to be produced by progressive melt intrusion of a dyke moving upward from a sill at 18 km depth in the midcrust of the volcanic rift zone (Fig. 20.11). A total of 545 events, recorded from the period July 6, 2007, to July 24, 2007, across the array, consisting of 27 seismometers, were selected for detailed study of earthquake foci clustering. The most remarkable feature of the seismicity was that

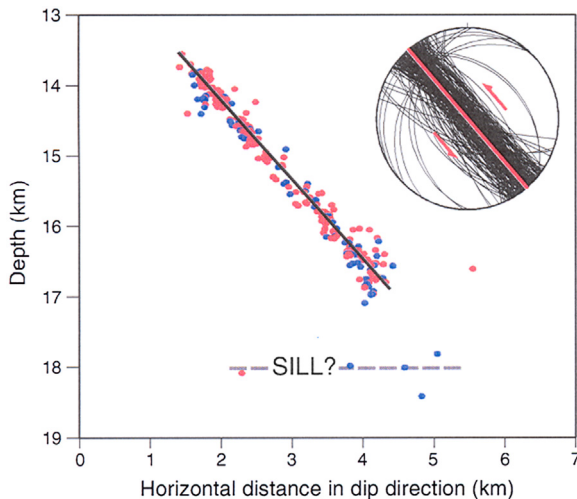


**FIGURE 20.11** Hypocenter locations and the possible position of the dyke viewed along strike of the dyke during the period July 6, 2007 to July 24, 2007 period of supposed melt injection at Upptyppingar. Dots show earthquakes with fault plane solutions: red are reverse faults and blue are normal faults. Inset shows fault planes from 229 double-couple reverse solutions viewed along the strike and projected onto an equal area projection orientated vertically and orthogonal to the dyke. Broken gray line shows the speculative location of a feeder sill. From White, R.S., Drew, J., Martens, H.R., Key, J., Soosalu, H., Jakobsdóttir, S.S., 2011. Dynamics of dyke intrusion in the mid-crust of Iceland. *Earth and Planetary Science Letters* 304, 300–312.

hypocenters shallower than 17.5 km, which were attributed by [White et al. \(2011\)](#) to intrusion along a dyke, lie on a single plane dipping at 50 degrees and striking at 74 degrees with a root mean square misfit of only 115 m. No surface intrusion was observed.

### 20.1.3.2 *Arabian Rift Zone Swarm*

From April 2009 to June 2009, a swarm of more than 30,000 earthquakes struck the Harrat Lunayyir lava field, situated at the northwestern end of the Saudi Arabian lava fields, east of the Red Sea. Harrat Lunayyir is one of the smallest Holocene lava fields in Saudi Arabia, but individual flow lobes radiate long distances from the center of the Harrat, and flows reach the Red Sea in two places. Harrat Lunayyir contains about 50 volcanic cones that were constructed over Precambrian crystalline rocks along an *N–S* axis. Lava flows are basaltic to basanitic in composition, and the Holocene flows are alkali olivine basalts. One of the cones may have erupted around the 10th century AD or earlier (GVP, 2011). The system of Saudi Arabian lava fields, situated along the western coast of the Arabian Peninsula, is connected with the Main Ethiopian Rift in northeastern Africa.

The seismic swarm began on April 18, 2009 ([Al-Amri and Al-Mogren, 2011](#)). The joint KSU and KACST seismic network, consisting of 21 seismic stations, was installed around the volcano, and 398 earthquakes with  $M > 3$  were located from April 29, 2009, to June 27, 2009, mainly from May 1 to May 21. Two epicentral areas were formed, to north and south of Harrat Lunayyir, leaving a free space beneath the volcano ([Fig. 20.12](#)). During the period from April 29 to May 12, seismicity developed to the south of the volcano. Three earthquakes at a depth between 13 and 15 km were recorded at the very beginning of this sequence on April 30 and May 1, and then the events clustered within the depth range between 2 and 7 km ([Fig. 20.13](#)). Earthquake frequency reached its maximum on May 12 and then sharply ceased. The maximum magnitude of the earthquake reached 3.7.

Beginning on May 12, seismic activity was recorded to the north from the volcano with a significant increase in the number and magnitude of earthquakes. The seismic activity reached its peak on May 19. A total of 12 earthquakes with  $mb \geq 4.0$  (USGS, 2011) were recorded on this day, including the  $M_w$  5.7 event. These seismic events clustered at a depth between 6 and 11 km, deeper than the initial southern group of earthquakes.

A northwest-trending 8-km-long surface fault rupture with 91 cm of offset was formed ([Pallister et al., 2010](#)) near the epicenter of the  $M_w$  5.7 earthquake ([Fig. 20.12](#)). The focal mechanism solutions published by Harvard University for three events, occurring on May 19 ([Fig. 20.12](#)), indicated the normal fault rupturing with the fault planes coinciding with the direction of the surface rupture and with the general direction of the surface structures of Harrat Lunayyir volcano (see inset in [Fig. 20.12](#)). Analysis of interferograms obtained by



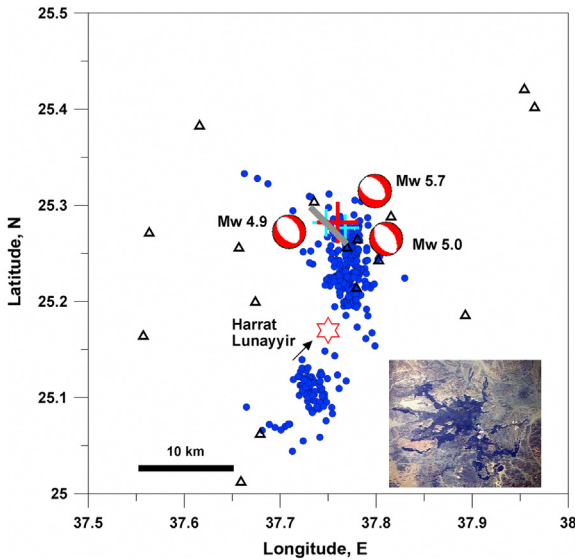


**FIGURE 20.12** Epicenters of earthquakes with  $M > 3.0$  located by the joint KSU and KACST seismic network at Harrat Lunayyir, Saudi Arabia, from April 2009 to June 2009. *Star* shows the volcano; *filled circles* show the earthquake epicenters. The epicenters of three large earthquakes with Harvard CMT solutions are shown with *crosses*; the *largest cross* indicates the position of the largest earthquake with magnitude  $M_w 5.7$ . The seismic stations are shown with *triangles*. The position of surface ruptures (according to [Pallister et al., 2010](#)) is shown with a *broad line*. *Inset* shows the satellite photo of Harrat Lunayyir (NASA Space Shuttle image STS26-41-61, 1988, <http://eol.jsc.nasa.gov/>). The catalog of earthquakes was taken from Al-Amri, A.M., Al-Mogren, S.M., 2011. *Seismo-Volcanic Investigation of the Current Activity in Harrat Lunayyir, Al-Madinah Al-Munawwarah Area. Final Report. King Saud University, Riyadh.*

Interferometric Synthetic Aperture Radar (InSAR) acquired for the period June 17 to June 28 ([Pallister et al., 2010](#)) showed that the deformation field at the volcano region was best modeled by the intrusion of a 10 km-long, NW-trending dyke, with a top at less than 2 km depth and volume of about  $0.13 \text{ km}^3$ . No surface intrusion was observed during 15 years after the seismic episode.

## 20.2 MODELING OF MAGMA ASCENT RESISTING

The data presented in [Section 20.1](#) showed that there is no difference in spatial-temporal distribution between the swarms of earthquakes observed in the cases of failed eruptions and the swarms of volcano-tectonic earthquakes, preceding the volcanic eruptions, which were described in Chapters 5–7. No difference with volcano-tectonic earthquake records was observed in the shape of the seismic signals associated with failed eruptions. Crustal deformation recorded for the majority of these cases allows to consider that this

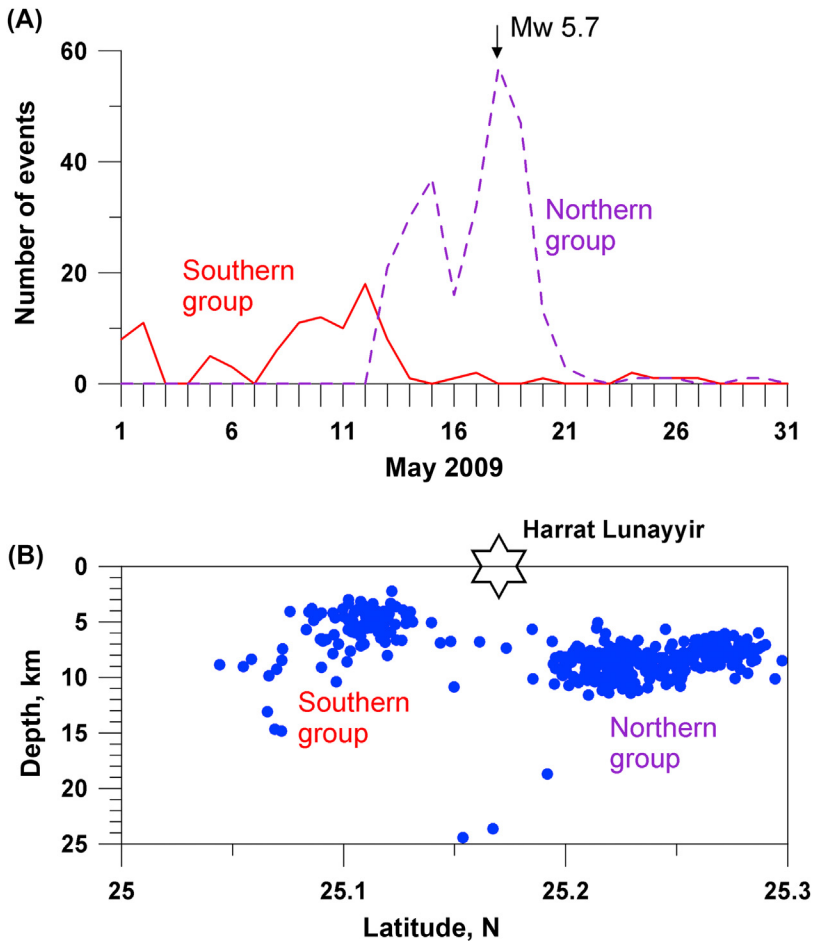


**FIGURE 20.13** Temporal development (A) and the meridional cross-section of earthquake swarm at Harrat Lunayyir (B). In (A), the *solid line* shows the variations in the number of the southern group of events and the *broken line* shows the variations in the number of the northern group of events. *Star* shows the volcano. *The catalog of earthquakes was taken from Al-Amri, A. M., Al-Mogren, S.M., 2011. Seismo-Volcanic Investigation of the Current Activity in Harrat Lunayyir, Al-Madinah Al-Munawwarah Area. Final Report. King Saud University, Riyadh.*

seismic activity beneath the volcanic constructions may be associated with any subsurface movement of magmatic matter that was arrested on its way to the surface. So the question is what determines whether or not a buoyant magma-filled crack actually reaches the Earth's surface or halts as a shallow intrusion.

Several mechanisms have been proposed for dyke propagation arrest, including magma solidification, reduction of magma chamber overpressure, barriers related to sharp gradients in mechanical properties, and density stratification.

Chen et al. (2011) noted that the dike arrest may be dominated by a single mechanism in some geological settings, or may be caused by the simultaneous action of multiple mechanisms. Magma solidification may dominate when a dike propagates slowly for a long period of time. Magma chamber overpressure reduction may be a major mechanism for dike arrest when the dike volume is significant compared with the volume of the magma chamber. Mechanical properties barriers and density stratification may act simultaneously when a dike propagates near the level of neutral buoyancy as the dike experiences marked changes in buoyancy force and mechanical resistance to propagation.

Some results of laboratory and numerical experiments are also presented in the following sections. They give the ideas about the process of dyke arrest below a volcano and, as a result, a case of failed eruption.

### 20.2.1 Experimental Study of the Ascent of a Fixed Magma Volume

Taisne et al. (2011) studied the downward propagation of a fixed volume of dense liquid through a lighter elastic–brittle medium made of gelatin. A fixed volume of dense liquid was injected in a small preexisting cut at the top of a large tank. The initial cut was deep enough to be unstable, such that the buoyancy force was sufficient to induce fracturing at the tip. The experiments were focused on cracks issuing from a source of small breadth  $B_0$  in the horizontal direction, such that the crack extended both horizontally and vertically.

Taisne et al. (2011) analyzed different phases of crack propagation. In the first phase, the crack grew larger in both breadth and length, driven by a lateral pressure gradient due to the liquid overpressure. It maintained an approximately circular shape. The propagation was characterized by continuous thinning of the nose region because of the loss of liquid to the growing tail region. The dyke eventually reached a limited width and extended only vertically. In the second phase, the crack maintained a constant breadth and extended only in the vertical direction with a parabola-shaped nose region. The final phase saw a rapid decrease in velocity and the arrest of propagation. In all the experiments, propagation was driven by buoyancy and resisted by forces associated with fracturing at the tip, elastic deformation, and viscous flow through the thin crack aperture. The cracks were characterized by the crack length in the vertical direction,  $L$ ; crack breadth in the horizontal direction,  $B$ ; and thickness of fluid within the crack,  $\delta$ . The relevant physical properties were the fracture toughness  $K_c$  and shear modulus  $G$  of the host solid and the buoyancy  $\Delta$  and viscosity  $\mu$  of the liquid.

Experimentally derived scaling laws allowed to estimate the dimensions and volume of the largest dyke, associated with a magma source at a depth  $D$ , that can stall below Earth's surface. Eq. (20.1) shows the dependence of the final crack length  $L_f$  on the fracture toughness  $K_c$  and the buoyancy  $\Delta\rho$

$$K_c = \gamma \Delta \rho g L_f^{3/2} \quad (20.1)$$

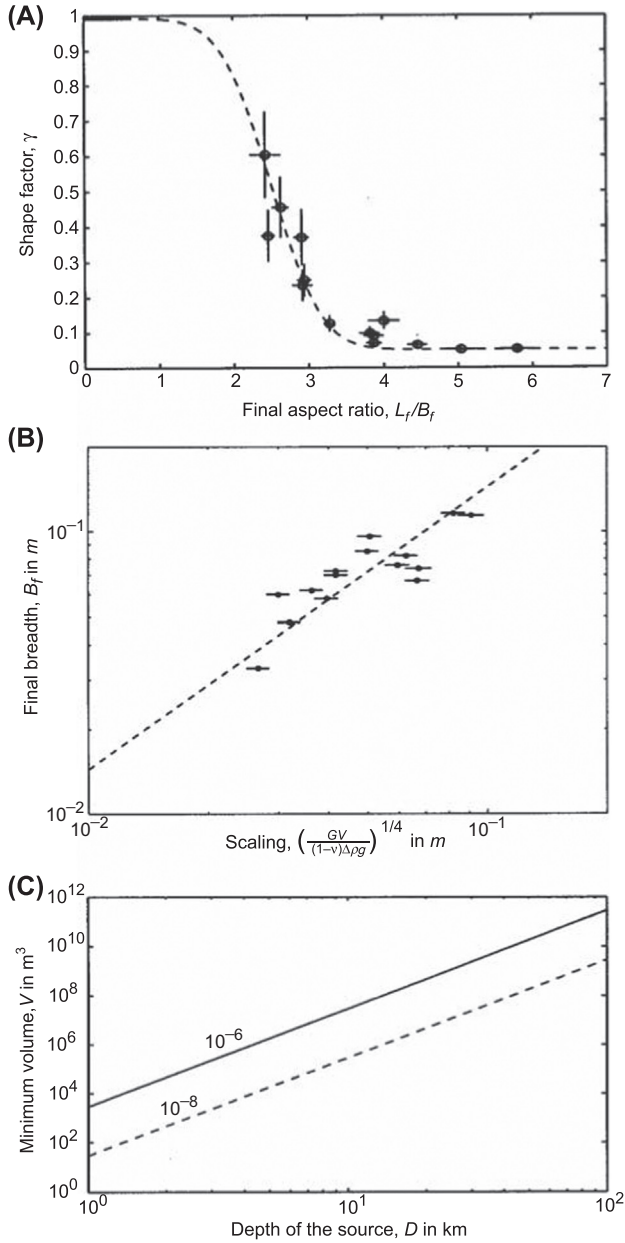


**FIGURE 20.14** Relationship between the parameters used for modeling of magma ascent to the surface. (A) Geometrical shape factor  $\gamma$  for the stress intensity factor of experimental cracks in gelatin as a function of the crack aspect ratio,  $L_f/B_f$ . In the limit of  $L_f/B_f \rightarrow 0$ , one retrieves the 2D crack theory, such that  $\gamma_0 = 1.12 \sqrt{\pi/2}$ , which is indicated by the thick bar at the left of the plot. The dashed curve was drawn through the data for illustration purposes. (B) Measurements of the final fracture breadth as a function of values predicted by a simple balance (Continued)

where  $\gamma$ , the shape coefficient, depends on the final aspect ratio,  $L_f/B_f$ , and  $g$  is the gravitational acceleration. Fig. 20.14A shows how  $\gamma$  varies as a function of the final crack aspect ratio.

The crack achieves its final breadth  $B_f$  during the first propagation phase. The balance between elastic and buoyancy forces leads to the following reasonable agreement with the experimental data (Fig. 20.14B):

$$B_f = \left( \frac{GV}{(1-\nu)\Delta\rho g} \right)^{1/4} \quad (20.2)$$

According to experimental results, the successful propagation of a dyke to the surface or its stalling is conditioned by the initial crack breadth, the final length of the crack, and the magma volume available. One condition for the stalling of a dyke below Earth's surface is that the final length of crack  $L_f \leq D$ . A stronger constraint on eruption comes from the magma volume available. It was found that the successful propagation of a dyke to the surface requires that the magma volume exceeds a threshold value. If the initial crack breadth is  $B_o$ , the threshold volume  $V_{\min}$  is given as follows:

$$V_{\min} = C_o B_o L_o \delta_o \quad (20.3)$$

where  $C_o$  is a proportionality coefficient of one order. The minimum magma volume that is required for an eruption is shown as a function of source depth  $D$  in Fig. 20.14C, which gives the final maximum vertical extension that a dyke can attain given a finite amount of magma. If the value found is smaller than the depth of the source, the magma cannot reach the surface, and hence there is no eruption.

### 20.2.2 Arrest of Propagating Dyke Due to Mechanical Barriers and Density Stratification in an Upper Crustal Horizon

The crust above the magma chamber is composed of layers with large and abrupt contrasts in stiffness. Because injection and arrest of dykes and sheets occur primarily in the layers above the source chamber of the volcano, the focus is on stiffness variation between those layers. In the uppermost parts of many active volcanic zones and central volcanoes, layers of soft sediments and pyroclastic rocks offer strong mechanical contrasts to the stiffer lava

between buoyancy and elastic stresses in a penny-shaped crack, which is appropriate for the first propagation phase (Eq. (20.2)). (C) Relationship between the volume of magma and the vertical length of a buoyancy-driven crack that stalls beneath Earth's surface (values for the physical properties are taken as  $\Delta\rho = 10$  and  $100 \text{ kg/m}^3$ ,  $G = 109$  and  $1010 \text{ Pa}$ , and  $\nu = 0.25$ ). This relationship can be understood as the minimum magma volume that is required for an eruption as a function of the depth of the magma source. The two curves were drawn for the two extreme values of ratio  $\Delta\rho g/G$  that are indicated. After Taisne, B., Tait, S., Jaupart, C., 2011. Conditions for the arrest of a vertical propagating dyke. *Bulletin of Volcanology* 73, 191–204.

flows. A soft, low-density layer may suppress the tensile stress associated with the tip of a dyke or sheet intrusion so much as to make it impossible for the intrusion to continue its extension.

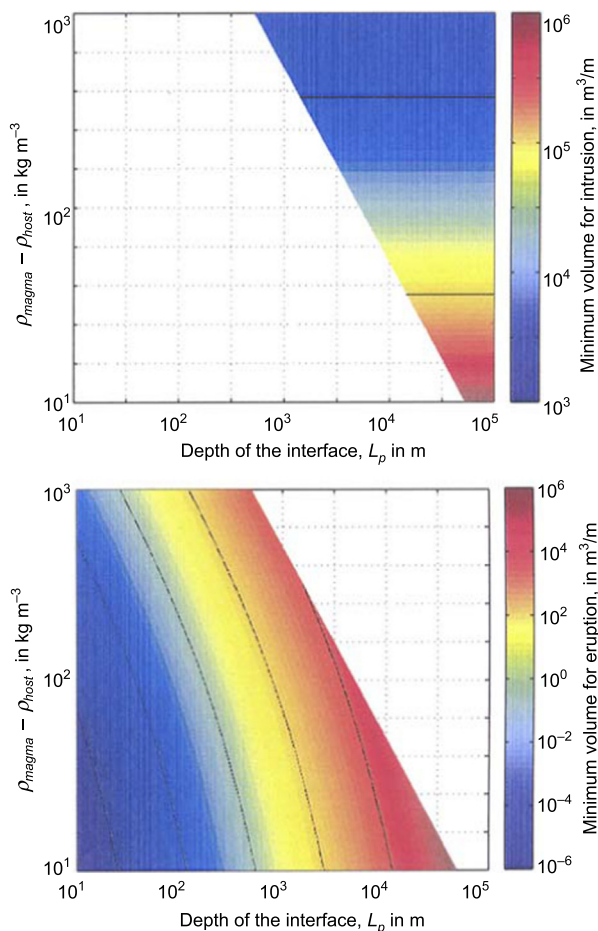


**FIGURE 20.15** Results of the numerical investigation of the minimum volume needed for a sill or for an eruption. *Top panel:* minimum volume needed for a sill as a function of the thickness of low-density rocks and density contrast between magma and host. Due to the negative buoyancy of magma, a dyke can only penetrate over a finite distance before reaching the rupture threshold for intrusion of a horizontal magma sheet. The blank region on the left is such that the interface is not deep enough to allow sufficient pressure build-up within the dyke. The two *black horizontal lines* correspond to volumes of 104 and 105  $\text{m}^3/\text{m}$ . *Bottom panel:* minimum magma volume for an eruption. The blank region on the right is such that the dense basement is too deep for the dyke to extend to the surface. Calculation for a value  $\Delta\rho_0 = 100 \text{ kg/m}^3$  of magma buoyancy below the interface. *Black curves* correspond to 10 – 4, 10 – 2, 100, 102, and 104  $\text{m}^3/\text{m}$ . From Taisne, B., Tait, S., Jaupart, C., 2011. Conditions for the arrest of a vertical propagating dyke. *Bulletin of Volcanology* 73, 191–204.

Gudmundsson (2002) used boundary element models to study the effects of a large variation in stiffness between layers on the stress field inside a volcano. He showed that the soft layer would act as a stress barrier where many sheets and dykes could become arrested. Gudmundsson (2002) noted also that many-layered rocks, such as in active central volcanoes and volcanic zones, contain sharp contacts between layers, and these contacts are often mechanically weak. When a horizontal discontinuity, such as a contact, a

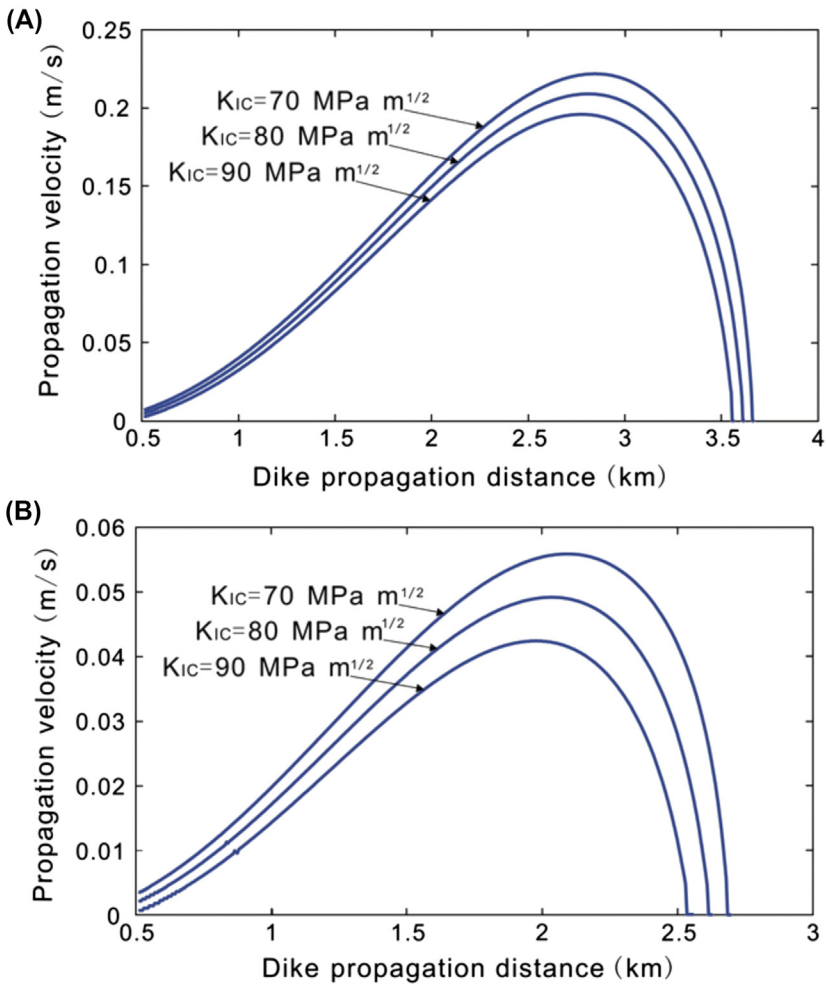


**FIGURE 20.16** Dike propagation velocity versus propagation length (A) Dike is 1 km below the level of neutral buoyancy (LNB); (B) dike is 1 km below the LNB. Collated after Chen, Z., Jin, Z.-H., Johnson, S.E., 2011. Transient dike propagation and arrest near the level of neutral buoyancy. *Journal of Volcanology and Geothermal Research* 203, 81–86.

joint, or a fault, is approached by the tip of a vertically propagating dyke, the discontinuity may open up and contribute to arresting the dyke.

A series of numerical calculations, performed by [Taisne et al. \(2011\)](#), showed that a low-density crustal layer can be an effective barrier in the path of vertical dyke propagation. When magma accumulates in a swollen nose region at the interface between the low-density layer and the dense basement, the magma overpressure is largest at this interface and increases with increasing penetration into the upper layer. It may become large enough to induce horizontal fractures in the dyke walls and lateral intrusion of a sill, which prevents eruption. This requires that the thickness of the low-density layer exceeds a threshold value that depends on the density contrast between magma and host rock. [Fig. 20.15](#) shows the obtained dependence of the minimum magma volume on the thickness of low-density rocks and density contrast between magma and host. If the magma volume is smaller than a threshold value, neither sill intrusion nor eruption is possible, and magma gets stored in a horizontal blade-shaped dyke straddling the interface.

[Chen et al. \(2011\)](#) numerically modeled the arrest of dyke propagating into the region above the level of neutral buoyancy (the depth where the host rock density becomes equal to the magma density) applying the finite difference/perturbation method. Numerical results showed that dyke propagation velocity initially increased with dyke length, reached a peak value when the dyke tip reached a position well above the level of neutral buoyancy, and then rapidly decreased to zero after the dyke propagates further into the region above the level of neutral buoyancy ([Fig. 20.16](#)). [Chen et al. \(2011\)](#) consider that these results indicate that density gradation is an effective mechanism for dyke arrest.

## 20.3 MONITORING OF THE SEISMIC ACTIVITY AT DORMANT VOLCANOES

As was shown in Chapters 5–8, the swarms of volcano-tectonic earthquakes occurring at active andesitic and dacitic volcanoes and at basaltic volcanoes are characterized by different features of their development. Therefore the author discussed separately the monitoring of seismic activity at the dormant volcanoes of these two groups of volcanic structures.

### 20.3.1 Monitoring of Andesitic and Dacitic Dormant Volcanoes

The problem of the reawakening of andesitic and dacitic dormant volcanoes was discussed by [Zobin and Jiménez \(2008\)](#) (see also Section 8.3). It was shown that is possible to discriminate the following three stages in the reawakening process: the preliminary stage that was characterized by a sharp increase in seismic activity near the volcano and continues till the first phreatic explosion; the stage between the first phreatic explosion and the large



explosion; and the postexplosion stage of dome building with a gradual decrease in the number of seismic events (Fig. 20.17B and C).

Seismic activity, associated with failed eruptions at andesitic and dacitic volcanoes discussed in Section 20.1.2, was also characterized by a sharp increase in seismic activity near the volcano. The first impulses of seismic activity recorded before an eruption and without eruption are similar. The only difference is the appearance of a phreatic explosion as a signal of forthcoming eruption. If the phreatic explosion was recorded at the moment of peak frequency of seismic events, the scheme of Fig. 20.17B may be taken as a more probable continuation of the seismovolcanic activity. If a phreatic explosion was recorded during continuous seismic activity, the scheme of Fig. 20.17C may be taken as a more probable continuation of the seismovolcanic activity.

For the majority of well-documented events, the duration of the first seismic impulse was no longer than 1 month. Therefore the final solution may be taken in as follows: If the first month of seismic activity was finished with a complete cease of seismic events, this seismic episode with a high level of probability was associated with a failed eruption. If the seismic activity continues for a longer time, we can only wait and continue monitoring till the end of the seismic episode.

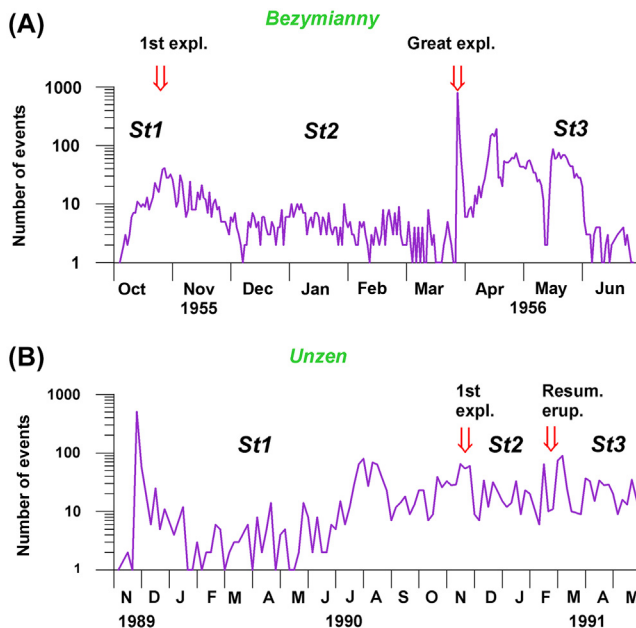


**FIGURE 20.17** Variations in the number of seismic events of swarms observed at Bezmianny volcano, Kamchatka (A) and Unzen volcano, Japan (B). St1, St2, and St3 indicate the stages of seismic activity. The data were taken from Tokarev, P.I., 1981. *Volcanic Earthquakes of Kamchatka*. Nauka, Moscow, 164 p. (in Russian), (Bezmianny) and the catalog prepared by Shimabara Earthquake and Volcano Observatory (Unzen).

It may be illustrated in the case of the 1983 Asacha swarm (see Fig. 20.9). This swarm reached the maximum frequency of events 7 days after the beginning of the activity. The high level of activity continued for 6 days, and then gradually decreased. The swarm was ceased after 24 days of activity. According to characteristic scenarios of the reawakening of andesitic and dacitic dormant volcanoes shown in Fig. 20.17, there are two ways of reawakening: a phreatic explosion may occur between the 7th and 13th days of maximum activity, and then the activity can develop according to the scenario shown in Fig. 20.17A. No phreatic explosion was recorded, and this scenario is rejected. Now, beginning from the 20th day of activity, the scenario of Fig. 20.17B may be applied for monitoring. But the swarm was over after 24 days of activity. Therefore about 1 month after the cessation of the swarm, the decision about failed eruption may be taken.

### 20.3.2 Monitoring of Basaltic Dormant Volcanoes

Activation of basaltic dormant volcanic structures may occur in three ways: the reawakening of a dormant volcano, the birth of a new volcano, and associated with dyke injections along rift zones. I know only one case of the reawakening of a basaltic volcano that occurred at Academia Nauk volcano, Kamchatka, on January 2, 1996. It was the first historic eruption of Academia Nauk volcano after 4800 years of repose (Belousov and Belousova, 2001). At the same time, the birth of new volcanoes (Kamchatka, Mexico, submarine events) and surface dyke injections along rift zones (Iceland, Afar) are frequent enough. Some characteristic features of the seismic activity associated with these eruptive events at dormant structures are explained below.

#### 20.3.2.1 Reawakening of Basaltic Volcano

The reawakening of Akademina Nauk volcano in Kamchatka after 4800 years of repose that occurred on January 2, 1996 (Belousov and Belousova, 2001; see also Section 2.3.1), may serve as an example of this type of activity. During 24 years before the eruption, the seismic activity developed at depths of 0–22 km, with the majority at depths between 5 and 15 km. The main preliminary swarms were recorded in 1978, 1992, and 1995. The maximum magnitude of 1972–95 events was  $M_s$  5.5 (Zobin et al., 1983).

A sharp increase in seismic activity was observed on January 1, 1996 (Fig. 20.18A). The process began with the 2-hour sequence of a total of 199 very small (magnitude about 0) events. The second sequence included 58 events of magnitude greater than 2. The majority of them occurred at a depth between 2 and 20 km, but nine events were recorded between 40 and 60 km of depth. This sequence continued for 4 hours and culminated in the  $M_w$  7.1 earthquake.

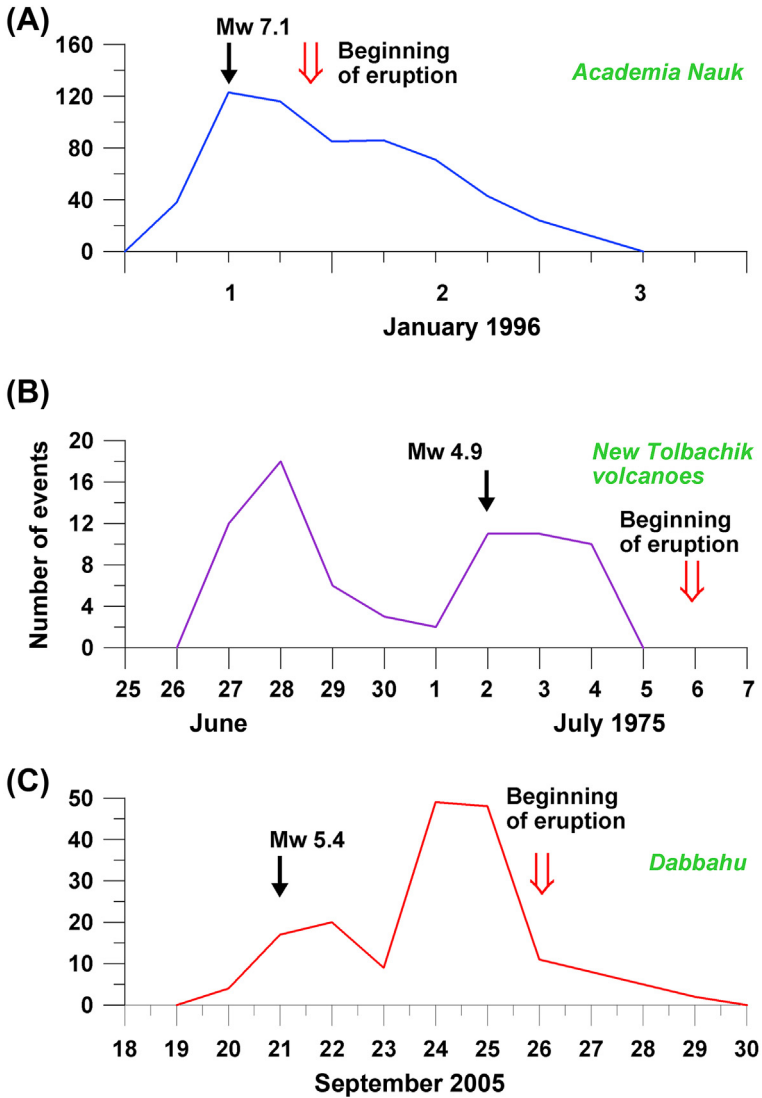


**FIGURE 20.18** Variations in the number of seismic events of swarms observed at Academia Nauk volcano, Kamchatka (A), New Tolbachik volcanoes, Kamchatka (B), and Dabbahu volcano, Ethiopia (C). The data were taken from the *Regional Catalog of Kamchatka Earthquake* (Academia Nauk and New Tolbachik volcanoes) and from the *U.S Geological Survey catalog* (Dabbahu).

The third sequence of seismic events occupied the area of the second sequence epicenters and spread also to the southwest along the regional fault. This sequence consisted of 261 events with magnitudes greater than 3 and continued for about 14 hours. The eruption of a new volcanic center in Karymskoye lake situated

within the Academia Nauk caldera occurred about 14 hours after the beginning of the aftershock sequence. The seismic activity continued then for about a month gradually decreasing (Zobin and Levina, 1998).

The main difference in the seismic activity just before the beginning of the 1996 eruption and the seismic activity of 1972–95 was in the deeper distribution of the foci of the January 1996 earthquake sequence and in the appearance of a large  $M_w$  7.1 earthquake.

### 20.3.2.2 *Birth of New Volcanoes*

The birth of new volcanoes at basaltic volcanic structures was recorded within the Michoacán-Guanajuato Volcanic Field, Mexico (Jorullo, 1759; Parícutin, 1943), within the Tolbachik Dol field of cinder cones, Kamchatka (New Tolbachik volcanoes, 1975), during submarine eruptions at Izu volcanic arc (Teishi-Knoll, 1989) and CoAxial Segment volcano (1993) within Juan de Fuca submarine ridge, among others. Seismic activity associated with some of them is described in Chapter 4.

Best-documented cases of the 1975 Tolbachik eruption (Section 5.3.1 and Fig. 20.17B) and the 1989 Teishi-knoll eruption (Section 5.5.1) allow to formulate the following characteristics of this type of activity: the seismic swarm continued for about 2 weeks, the largest earthquake occurred during the second peak in the seismic activity, and the eruption of new volcanoes began in a few hours after ceasing or at low background level of the final stage of seismic activity.

### 20.3.2.3 *Dyke Injections along Rift Zones*

The best-documented case was recorded during the 2005 dyke injection along the Dabbahu-Manda Harraro rift segment (Ethiopia) of the Afar triple junction zone (see Section 5.3.2). It was preceded by at least two preliminary earthquake swarms recorded in 1978 and 1989 (Ayele et al., 2009).

The seismic crisis began on September 4, 2005. Activity subsided but then restarted on September 20. Fig. 20.18C shows the development of this main part of the swarm according to the USGS catalog. It continued for about 2 weeks. The largest earthquake occurred during the first peak in the seismic activity. The eruption of the new volcanic center began when the seismic activity gradually decreased during its final stage.

### 20.3.2.4 *Possible Scenarios for Monitoring of Basaltic Dormant Volcanoes*

Data presented earlier allow us to formulate some regularity in the development of seismic activity associated with basaltic eruptions. Periods of seismic activity may be separated into three stages:

Stage 1, preliminary: A sequence of seismic swarms without eruptions; may continue for a few years.

Stage 2, preceding: Seismic swarms occurring just before the eruption; may continue during time intervals from a few hours to a few weeks.

Stage 3, continuous: A sequence of seismic swarms occurring after the initial eruption with the formation of new volcanic centers; may continue for a few years.

Table 20.1 shows the presence of these stages (or part of them) for different types of eruptions.

Now we can discuss a possible strategy for monitoring basaltic dormant volcanoes. We have a seismic swarm in a volcanic zone where no seismic activity was recorded for a long time. According to the characteristics of the seismic activity described earlier and shown in Fig. 20.18 and Table 20.1, it may be a preliminary swarm before the reawakening of a dormant volcano, a swarm preceding the birth of a new volcano or dyke injection, or the earthquake swarm associated with the failed eruption.

Our strategy may be illustrated in the case of the 2009 Saudi Arabian swarm at Harrat Lunayyir (Figs. 20.12 and 20.13). The first option may be that this seismic swarm is preliminary, representing stage 1 from Table 20.1. In this case, the probable scenario would be a process similar to the 2005 dyke injections along rift zones in Afar (Fig. 20.18C). It is a structure very similar to the Saudi Arabian volcanic rifts, having a preliminary seismotectonic crisis between 1978 and 1989. In this case, the 2009 Harrat Lunayyir episode would have a continuation of the seismic activity (stages 2 and 3) within the Saudi Arabian harraat adjacent to Harrat Lunayyir and accompanied by the surface dyke intrusions with the formation of new eruptive centers. Monitoring is needed for the following few years.

The next probable scenario, having also stage 1, may be the reawakening of the old Harrat Lunayyir volcano. This process, observed at Akademia Nauk volcano (Fig. 20.18A), was characterized by three preliminary low-magnitude (up to  $M_s$  5.5) and shallow earthquake swarms during 8 years before the volcano reawakening. The appearance of new earthquake swarms within the Harrat Lunayyir lava field may be a signal of this type of scenario. Monitoring is needed for the following few years.

**TABLE 20.1** Characteristic Durations of Three Stages in Seismic Activity Associated With Basaltic Eruptions.

| Type of Eruption           | Stage 1    | Stage 2 (h) | Stage 3    |
|----------------------------|------------|-------------|------------|
| Eruption at active volcano |            | 0–120       |            |
| Newborn volcano            |            | 400–700     | 2–3 months |
| Reawakening volcano        | Some years | 20–30       |            |
| Rift intrusion             | Some years | 400–600     | Some years |

The birth of a new volcano at Harrat Lunayyir is less probable. This activity (Fig. 20.18B) would be characterized by a 2–3-week earthquake swarm with the maximum magnitude earthquakes of about 5–5.5 and culminated in the birth of a new volcano. From this point, the 2009 swarm had finished already without the appearance of a new volcanic structure within the Harrat Lunayyir lava field, and the process is over.

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## RELEVANT WEBSITES

- GVP, Global Volcanic Program, Smithsonian Institution, <http://www.volcano.si.edu/world/volcano>.
- Harvard CMT, <http://www.globalcmt.org/CMTsearch.html>.
- USGS, U.S. Geological Survey. <http://earthquake.usgs.gov/>.

## Chapter 21

# The Seismic Signals Associated With the Natural Seismicity of Geothermal Structures within Volcanic Environment

Volcanic environment is a region that includes a variety of volcanic structures such as active volcanoes, dormant volcanoes, and geothermal systems. The activity of geothermal systems within volcanic environments represents the process that may be named as a mild eruption. The seismicity of these mild eruptions is a subject of the recent chapter. The author presents a review of the seismic signals associated with the natural manifestations of geothermal activity within the volcanic environment. The origin of these seismic signals is supposed to be related to the movement of hot steam and water bubbles within the surface-heated soil and/or in the water and steam columns. The induced seismicity associated with the manifestations of geothermal activity is not discussed.

### 21.1 GENERAL DESCRIPTION OF GEOTHERMAL SYSTEMS

Geothermal system is a natural heat transfer within a confined volume of the Earth's crust where heat is transported from a heat source to a heat sink, usually the free surface. The transport of heat to the surface in volcanic regions occurs by a convective process associated with the rise of hot magma. The heat is passed from the cooling magma in the chamber to the surrounding country rocks. Where these chambers or the rock above the chamber contain groundwater bodies, the water expands and rises buoyantly, to be replaced by cold water flowing in from the sides. This process establishes a circular geothermal system (Hochstein and Browne, 2000; McCall, 2005).

Many surface manifestations of heat discharge in volcanic regions may be classified as diffusive, continuous, or intermittent. Diffusive heat transfer by thermal conduction usually results in elevated near-surface temperatures and can be recognized mainly by temperature surveys in shallow boreholes. A mud volcano is one of the special surface manifestations in the regime of



diffusive heat discharge. The most common manifestations of continuous heat discharge are through convection in hot water springs. Another example is a hot steam at the saturation temperature through fumaroles. Geysers are spectacular manifestations discharging intermittently a mixture of boiling water, gas, and steam (Hochstein and Browne, 2000). Discharge of hot water and steam to the surface may produce seismic signals.

### 21.1.1 Position of Geothermal Systems Within Volcanic Environment

Figure 21.1 shows an example of the position of geothermal systems within the Kamchatkan volcanic environment. The Kamchatka Peninsula is situated at the northern end of the Kurile-Kamchatka island arc and is characterized by a high level of seismic and volcanic activity as the result of interaction between the continental North American plate and the subducting oceanic Pacific plate. The volcanic environment of Kamchatka is separated into two volcanic belts: Eastern (E\_K) and Central (C\_K). The Central belt is represented by only one active volcano but a great number of dormant volcanoes and hot springs, and two high-temperature geothermal systems. The Eastern belt is more active and consists of 28 active volcanoes, 9 high-temperature geothermal systems (one of them is the famous Geyser Valley), and numerous dormant volcanoes and hot springs. Both volcanic belts were formed during the Quaternary period (Leonov, 2001).

Figure 21.1 demonstrates that all high-temperature geothermal systems in Kamchatka are situated along the chains of active volcanoes. Hot fumarolic fields with temperatures higher than 400°C have been acting for centuries in the craters of Tolbachik, Mutnovsky, Avachinsky, Zhupanovsky, and Gorely. Acid and hot lakes are located in the craters of Maly Semiachik, Mutnovsky, and Gorely volcanoes. Temperatures as high as 220°C and 240°C have been measured in strong flank fumaroles of the Koryaksky and Kizimen volcanoes. Boiling point fumarolic fields are located at volcanoes of the Eastern volcanic belt such as Komarov, Bolshoy Semyachik, Burlyaschii, Kikhpinych, Dzenzur, Zheltovsky, Ksudach, Kambalny, and Koshelevsky. A fumarolic field exists at the slope of the Ichinsky volcano, which is the only active volcano in the Central volcanic belt. Four high-temperature geothermal systems situated within the Eastern volcanic belt have been drilled, and the first geothermal plant in Pauzhetka in southern Kamchatka has been in operation since the end of 1966. Mutnovsky geothermal field provides now >60 MW of electric power (Taran, 2009).

### 21.1.2 Structure of Geothermal Systems

Figure 21.2 shows two conceptual models of the structure of geothermal systems (Hochstein and Browne, 2000). The general sketch of the deep structure of the geothermal system (Fig. 21.2A) displays the main elements of the system. Heat

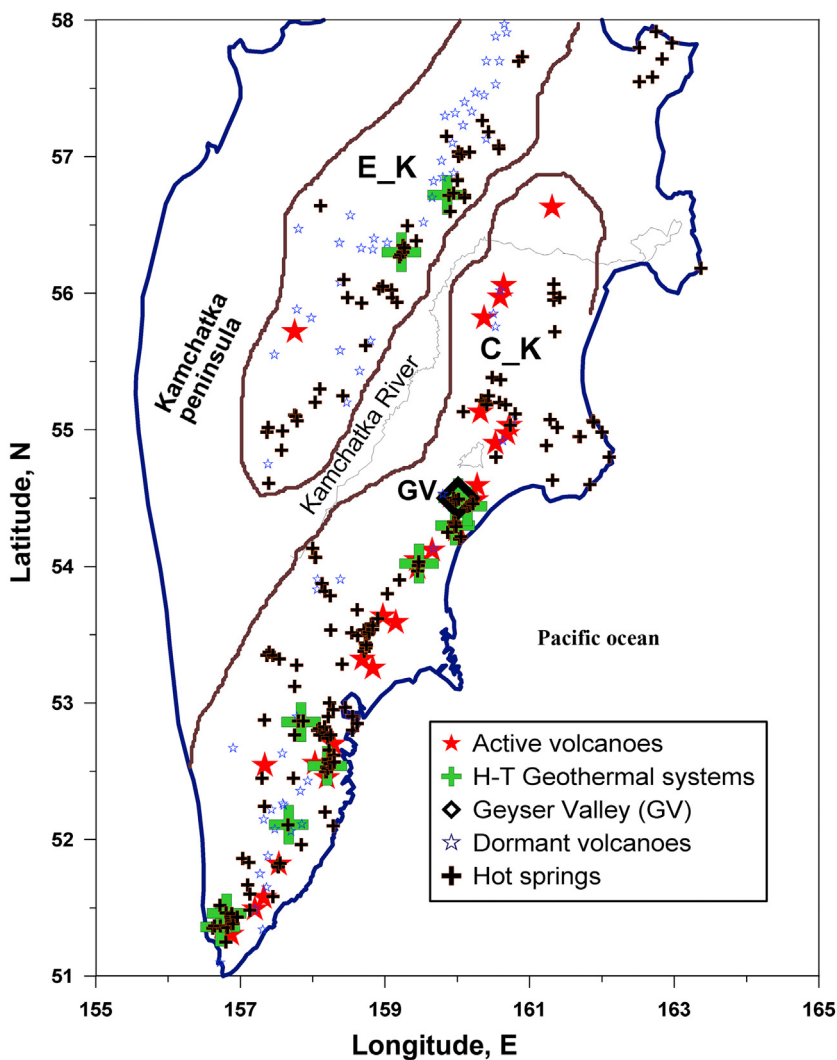


**FIGURE 21.1** Volcanic environment of Kamchatka Peninsula, Russia. There are shown active and dormant volcanoes, hot springs and high-temperature (H-T) geothermal systems, and Geyser Valley. E\_K, Eastern Volcanic Belt; C\_K, Central Volcanic Belt. *Catalog of active and dormant volcanoes* is taken from Simkin, T., Siebert, L., 1994. *Volcanoes of the World*, second ed. Geoscience Press, Tucson. 349 p. *Catalog of hot springs* is taken from Kiryukhin, A.V., Kiryukhin, V.A., Manukhin, Yu.F., 2010. *Hydrogeology of Volcanogens*. Nauka, S-Petersburg, 395 p. (in Russian). *Catalog of high-temperature geothermal systems* is taken from Sugrobov, V. M., Yanovsky, F.A., 2001. *Geothermal fields of Kamchatka. Heat losses by volcanoes and hydrotherms*. In: Fedotov, S.A. (Ed.), *Active Volcanoes of Kamchatka*. Nauka, Moscow, pp. 58–66 (in Russian).

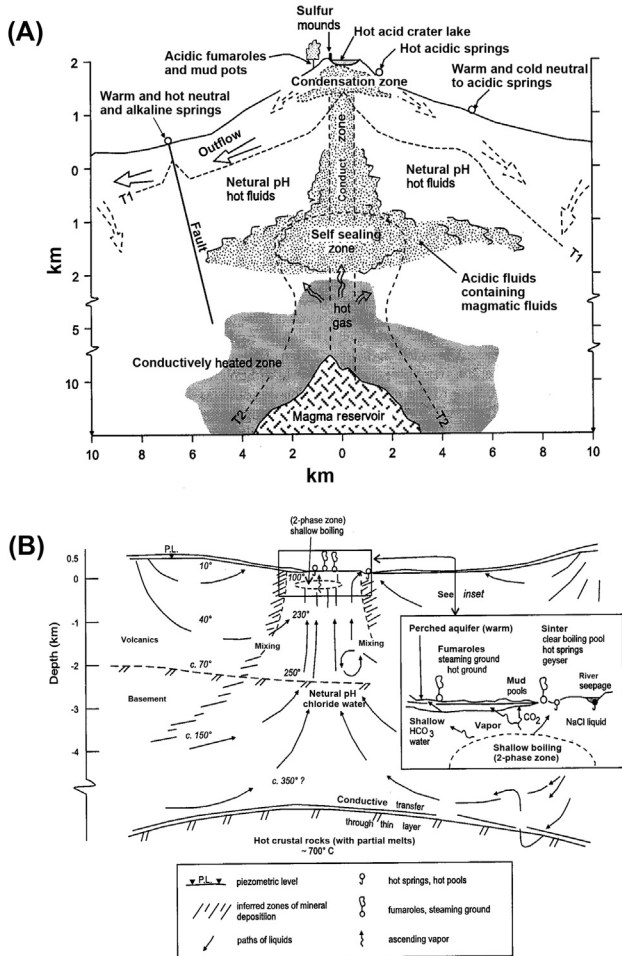


**FIGURE 21.2** Conceptual models of the deep structure of geothermal systems and their surface manifestations (A) A general sketch. Broken lines with T1 and T2 represent inferred isotherms for 150°C and 350°C, respectively. (B) Conceptual model of a liquid-dominated geothermal system. After Hochstein, M.P., Browne, P.R.L., 2000. Surface manifestations of geothermal systems with volcanic heat sources. In: Sigurdsson, H. (Ed.), *Encyclopedia of Volcanoes*. Academic Press, San Diego. pp. 835–855.

transfer from the magma reservoir through a conductively heated zone is realized by convective processes. The convecting heat transfer medium is dominantly infiltrated surface water. The fluids that transfer most heat are magma and magmatic gases. The surface manifestations are spatially zoned with respect to their volcanic center and are represented by solfataras, fumaroles, hot acid lakes, acid hot springs, and acid streams. At lower elevations, minor thermal springs may discharge neutral pH, chloride, or sometimes bicarbonate-chloride waters.

The special case of a liquid-dominated geothermal system is shown in Fig. 21.2B. The heat source is an extensive layer of hot crustal rocks that contains some partial melts and host intrusions. Because of their high recharge rate, the hot fluids can ascend close to or reach the surface. This thermal regime favors geothermal eruptions. The surface manifestations of this type of system are boiling hot springs and geysers.

## 21.2 NATURAL SEISMICITY ASSOCIATED WITH HEAT DISCHARGE WITHIN GEOTHERMAL SYSTEMS

Natural seismicity associated with three types of heat discharge within the geothermal systems (diffusive heat discharge, continuous heat discharge, and intermittent heat discharge) is discussed.

### 21.2.1 Diffusive Heat Discharge (Mud Volcano)

As was mentioned previously, the mud volcanoes represent the surface manifestations in the regions of diffusive heat discharge. There are more than 1500 mud volcanoes known all over the world that are normally in a dormant stage due to the short duration of eruptions. Their dormant stage activity is often characterized by vigorous seepage of water, gas, and petroleum. The only case of the birth of a mud volcano recorded seismically was the Lusi mud volcano in Indonesia. The source of the fluids and the fluid–rock interactions within the mud volcano conduit remain poorly understood (Mazzini et al., 2009).

#### 21.2.1.1 Newborn Lusi Mud Volcano, Indonesia

On May 29, 2006, at around 05 hours (May 28, 22 hours GMT, local time—7 hours) in the Sidoarjo district in East Java, Indonesia witnessed the sudden birth of numerous gas and mud vents on the ground. These eruption sites extended over a distance of more than a kilometer forming an NE–SW alignment. The eruptive bursts of hot water and steam were dramatic with a distinct geyser-like cycle of active and passive periods. Within a few days, a prominent crater formed, and the initial vents were quickly covered by the large volumes of erupted boiling mud. The flow rates from the main crater reached 180,000 m<sup>3</sup>/day. The largest crater of the new mud volcano was named Lusi [acronym from LUMPUR (mud) and Sidoarjo (the district name)] (Mazzini et al., 2007; Sawolo et al., 2009). The samplings of the fluids and the solids erupted by Lusi showed the presence of mud breccia and significant amounts of methane, typical of mud volcanoes worldwide. Preliminary analyses of erupted material indicated that the main source of clay and water was the overpressured units located at ~1500–1800 m that were rapidly buried and undercompacted (Mazzini et al., 2007).

Figure 21.3 shows the seismic record obtained on May 28 and May 29 at Tretes BMG station situated at a distance of 15 km from the mud volcano.

**Explosion of mud volcano LUSI, East Java (29 May 2006, 05h LT; 22h 28 May, GMT)**  
**Seismic station TRETES (epicentral distance 15 km)**

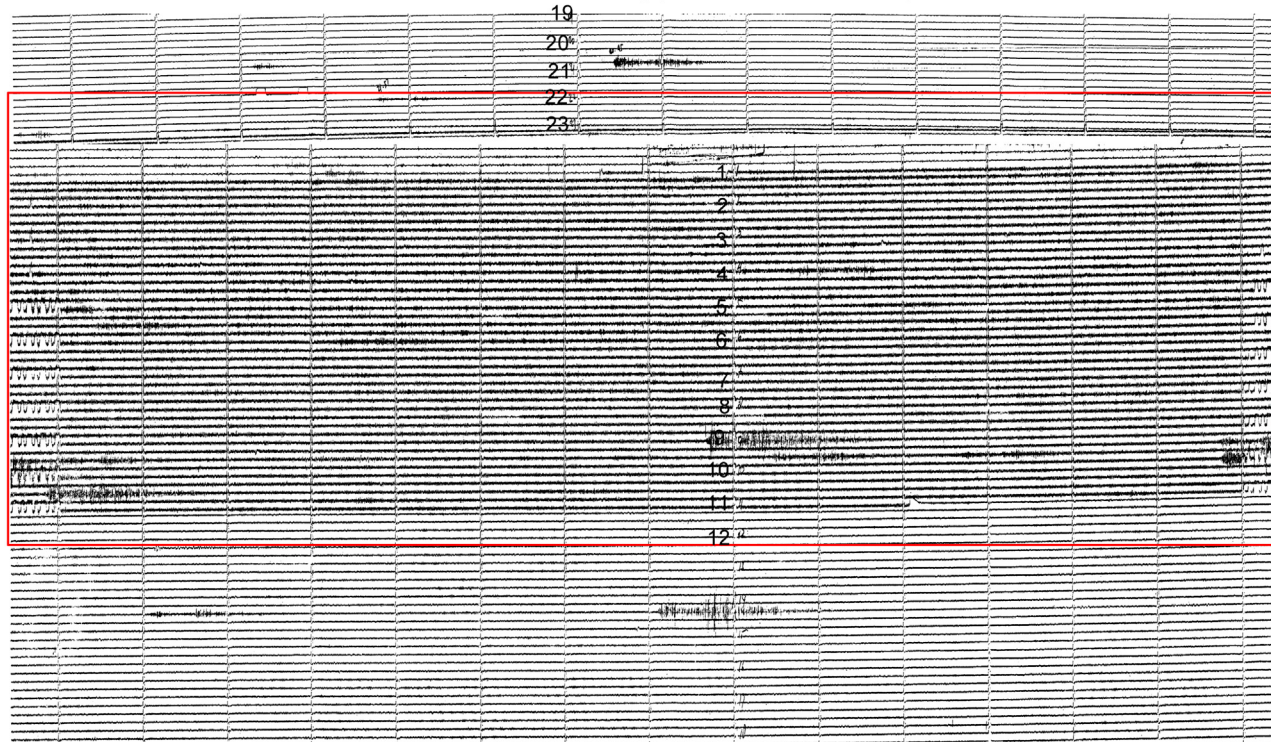


**FIGURE 21.3** Short-period seismic record, associated with the eruption of Lusi mud volcano, Indonesia, obtained on March 28, 2006, and May 29, 2006, at the seismic station Tretes situated at a distance of 15 km from the mud volcano. The 10-h interval from 1 to 11 h of May 29 (GMT) characterized by high-frequency, low-amplitude continuous seismic signals, recorded in about 3 h after the beginning of the eruption, is shown in rectangle. *Courtesy of Dr. Bambang P. Istadi, Energi Mega Persada.*

It is clear seen the 10-hour interval from 1 to 11 hours of May 29 (GMT) was characterized by high-frequency, low-amplitude continuous seismic signals recorded about 3 hours after the beginning of the eruption. A few high-frequency, high-amplitude signals of duration up to 2 minutes were recorded during this period and later. The largest of them were recorded at 08:59 and 14:14 on May 29 (GMT).

Fourier spectra, calculated for 30-second time windows of low-amplitude continuous records, high-amplitude events, and background seismic records (Fig. 21.4A) are shown in Fig. 21.4B. High-frequency events were characterized by the dominant frequency bands between 15–23 and 23–40 Hz. The dominant frequencies of low-amplitude continuous signals were between 5 and 20 Hz. For comparison, is shown the spectrum of the background signal having significantly lower spectral amplitude with a peak frequency at 12 Hz. At higher frequencies, well expressed for the continuous signal, the spectral amplitude gradually decreased.

#### 21.2.1.2 *Dormant Dashgil Mud Volcano, Azerbaijan*

Dashgil mud volcano is a 2-km-wide dome located in Azerbaijan near the western coast of the Caspian Sea. The summit area of the Dashgil volcano (Fig. 21.5) presents the typical structure of a mud volcano summit caldera. It is one of the most active mud domes of the Earth. Beyond its paroxysmal activity, it also represents a good example of a mud volcano with high seep activity in the dormant period. Gas, mud, and water seepage occur at a number of different structures. In the southwestern sector of the summit area, a north–south-oriented field of active gryphons and mud cones exist. A deep bubbling sound and a faint oil smell are recorded inside at least some of the cones. Toward the eastern edge of the caldera, two lakes of low-viscosity mud (salsas) exist. In both, gas bubbles rise from the bottom of the lakes, which reach maximum depths of about 10 m (salsa A) and 8 m (salsa B), respectively (Kopf et al., 2009).

Monitoring the activity of the Dashgil mud volcano in its “dormant” phase was carried out by using a portable short-period seismometer Tromino (Albarelli et al., 2012). Three sessions of seismic measurements were carried out in the crater area in June 2006 and July 2007 (sites I–III in Fig. 21.5A). Each session lasted 20 minutes and a sampling rate of 128 Hz was adopted. The two time series I (near crater) and II (near the lake of low-viscosity mud, salsa A) appear to be quite different to visual inspection (Fig. 21.6). Ground motion amplitude at the mud cone (site I) is about one order of magnitude larger than at the site II near salsa A. This is the effect of energy bursts (see their shape in the inset A1 in Fig. 21.6) that are present in the first series and absent in the second series. The 2007 measurement session was performed near mud cones (site III, Fig. 21.5). The energy bursts were clearly observed again (see the inset C1 in Fig. 21.6) confirming the possible presence of phenomena reflecting persistent activity in the summit area of mud volcano.



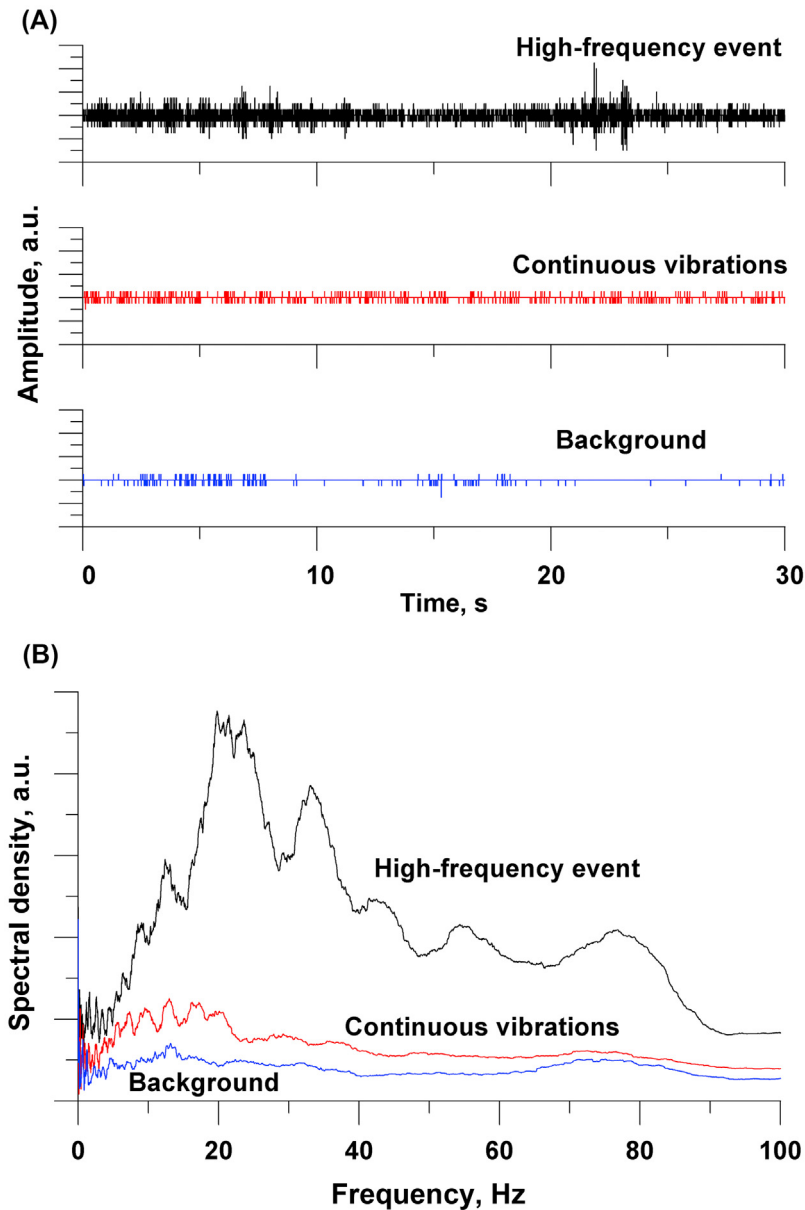


**FIGURE 21.4** Seismic records (A) and Fourier spectra (B) of 200 Hz sampling data associated with the eruption of Lusi mud volcano, Indonesia on May 29, 2006. Short-period seismic station Tretes, 15 km from the mud volcano. In (A), the upper line shows the high-frequency event recorded at 09 h, the middle line shows the continuous vibrations recorded at 10 h, and the lower line shows the background vibrations recorded at 12 h (see Fig. 20.3). The unit (a.u.) means arbitrary units. *Courtesy of Dr. Bambang P. Istadi, Energi Mega Persada.*

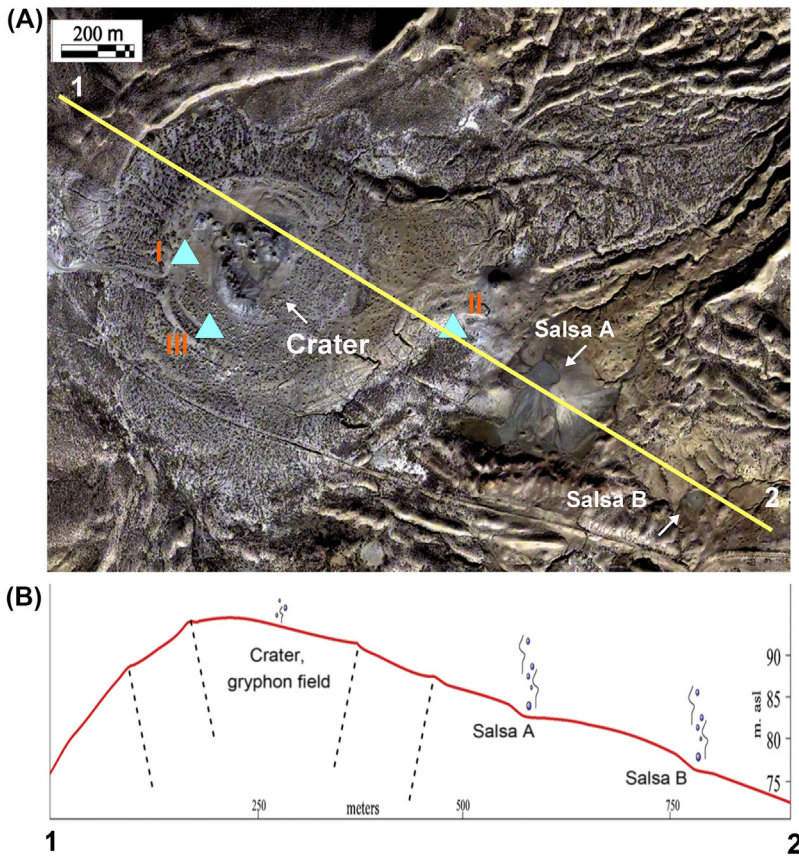


**FIGURE 21.5** Map of the Dashgil mud volcano crater field (A). Triangles indicate positions of ambient vibration measurements carried out in 2006 (sites I and II) and 2007 (site III). Positions of the crater and two lakes of low-viscosity mud (salsas A and B) are indicated with arrows. The 1–2 line indicates the trace of the section in the inset at the bottom of (B). Modified from Mazzini, A., Svensen, H., Planke, S., Guliyev, I., Akhmanov, G.G., Fallik, T., Banks, D., 2009. *When mud volcanoes sleep: insights from seep geochemistry at the Dashgil mud volcano, Azerbaijan*. *Marine and Petroleum Geology* 26, 1704–1715 and Albarello, D., Palo, M., Martinelli, G., 2012. *Monitoring methane emission of mud volcanoes by seismic tremor measurements: a pilot study*. *Natural Hazards and Earth System Sciences* 12, 3617–3629.

The shape of the energy burst signals recorded at Dashgil dormant mud volcano may be comparable with the shape of the high-amplitude, high-frequency signals recorded at the newborn Lusi mud volcano (see Fig. 21.4). At the same time, they are quite different in spectral content, having the dominant frequencies within the interval between 3 and 6.5 Hz for Dashgil volcano and within the interval between 15 and 40 Hz for Lusi volcano.



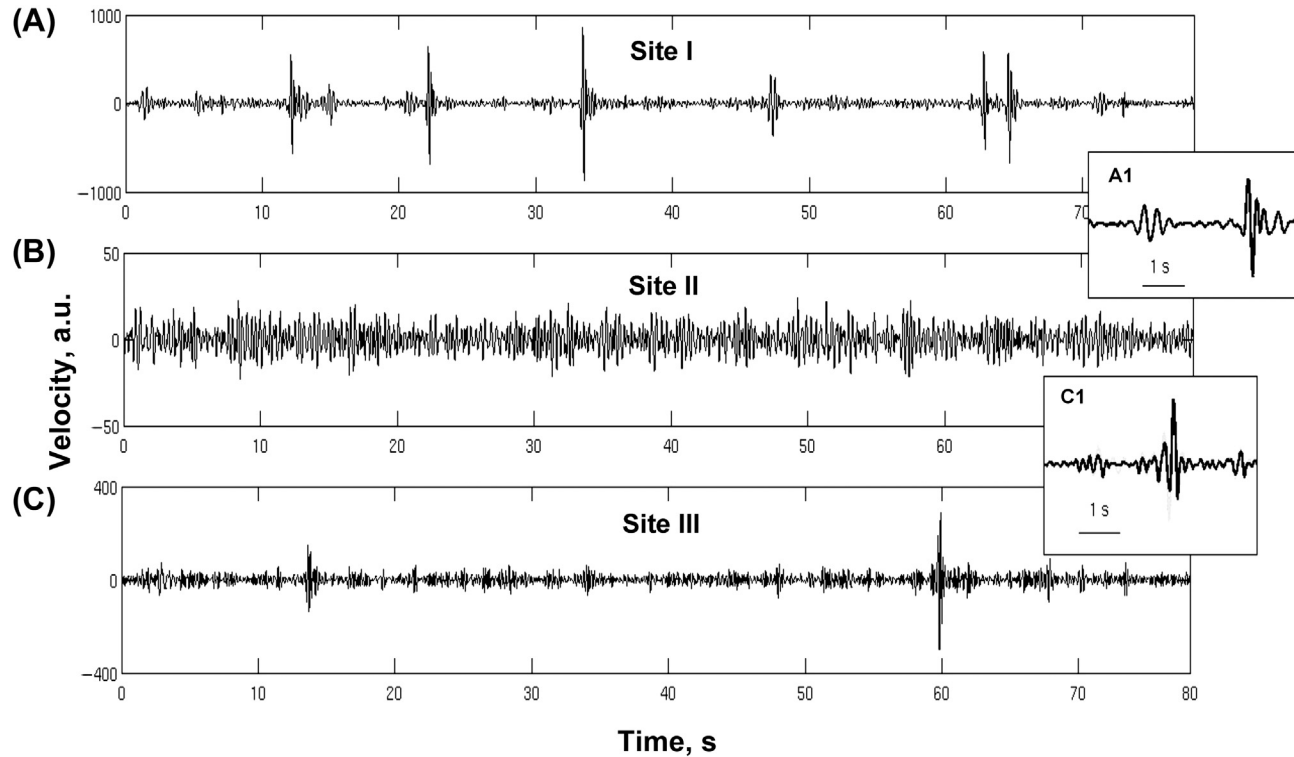


**FIGURE 21.6** Samples of the time series of seismic signals (vertical component) registered at the Dashgil mud volcano. (A) Series I measured on June 2006 (in the small inset aa, a magnification of the energy bursts observed in the series is shown); (B) Series II (June 2006); (C) Series III (July 2007); magnification of an energy burst for this last series is reported in the inset cc. *From Albarello, D., Palo, M., Martinelli, G., 2012. Monitoring methane emission of mud volcanoes by seismic tremor measurements: a pilot study. Natural Hazards and Earth System Sciences 12, 3617–3629.*

### 21.2.2 Continuous Heat Discharge

This type of heat discharge is represented by hot water springs forming geothermal systems. Geothermal systems are zones where significant heat and mass redistribution between a magma reservoir and the ground surface takes place (Norton, 1984). Geothermal systems usually appear on active volcanoes in response to the geothermal fluid circulation (i.e., magmatic gases, groundwater, meteoric waters, and seawater).

#### 21.2.2.1 Dallol, Ethiopia

The Dallol geothermal system is located in the Danakil depression, in the Afar Province, Ethiopia. Dallol has a superficial volcanic system associated with post-eruptive activity; its last recorded eruption was in 1926. The salt lake, situated within the depression, represents an important source of volcanic heat that survives, causing the ascent of hot water through the layers of halite and anhydrite. The minerals are dissolved and deposited over the basaltic lava flows, originating fumaroles and formations called hornitos, small rootless spatter cones that often form on the surface of a basaltic lava flow (Carniel et al., 2010).

The broadband and short-period seismic instruments that were installed near hornitos of Dallol during December 2003 by Carniel et al. (2010) showed the presence of two regimes of the continuous geothermal tremor with predominant groups of frequencies around 0.5–2, 2–12, and 8 Hz (Fig. 21.7). The first regime was characterized by low-amplitude continuous signal (Fig. 21.7A) while the second type of regime was characterized by the presence of repeating strong pulses with duration of about 4.5–5 seconds, repeated every 2–14 seconds (Fig. 21.7B).

The continuous data indicate variable shallow processes most likely related with variations in temperature and degassing processes within the shallow geothermal system. The depth of the heat source of perturbations beneath the hornitos was estimated as very superficial, less than 100 m (Carniel et al., 2010).

#### 21.2.2.2 Uzon, Kamchatka

The Uzon caldera hosts a large geothermal field and numerous hot and cold lakes. Present geothermal activity is concentrated in a  $0.2\text{--}0.3 \times 5$  km zone filled with boiling springs, gas–steam jets, mudpots, small mud volcanoes, hot lakes, and springs. The latest phreatic eruption took place in 1989 and created a 14-m wide crater. The Uzon caldera hosts a few Holocene volcanic vents (Leonov et al., 2001).

Figure 21.8 shows short-period seismic records obtained in the Uzon geothermal system by Kugaenko et al. (2009) from November 2007 to December 2007. The geothermal background tremor was characterized by

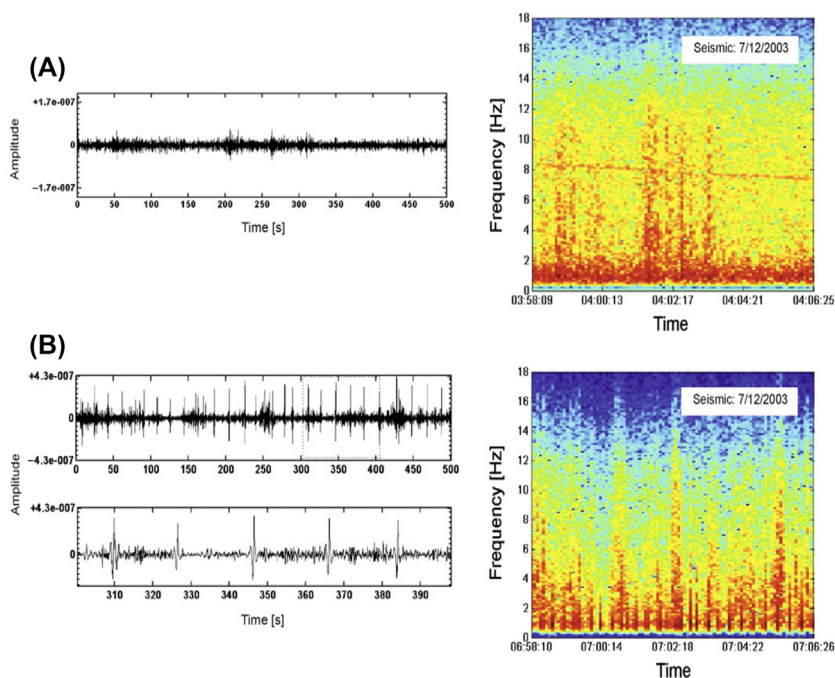


**FIGURE 21.7** Two types of geothermal tremor (short-period, 100 Hz sampling, seismic signals and corresponding spectrograms) recorded near hornitos of Dallol, Ethiopia, geothermal system on December 7, 2003. (A) Low-amplitude tremor and (B) high-amplitude tremor with strong pulses. In (B), the interval between 300 and 400 s is shown in detail. After Carniel, R., Muñoz Jolis, E., Jones, J., 2010. A geophysical multi-parametric analysis of hydrothermal activity at Dallol, Etiopía. *Journal of African Earth Sciences* 58, 812–819.

the peak frequency at 3.5 Hz (Fig. 21.8A and B). There were recorded also 5–7-second-long signals (Fig. 21.8C and D) characterized by the frequency band of 3–9 Hz. Kugaenko et al. (2009) noted the repetitive nature of these signals.

### 21.2.2.3 Satsuma-Iwojima, Japan

Satsuma-Iwojima is a volcanic island located on the northwestern rim of the submarine Kikai caldera, whose last large eruption occurred about 6300 years ago. The small ash eruptions were recorded during 1988–2004 (GVP, 2012). The eastern half of the island is occupied by a rhyolitic cone called Iwodake, which has an active crater at its summit of 400 m in diameter. At the bottom of the crater, there is a gradually expanding fumarolic vent, which continuously and steadily emits high-temperature volcanic gas at nearly 900°C. In addition to the central vent, there are many small, active fumarolic vents in and around the summit area of Iwodake (Ohminato, 2006).

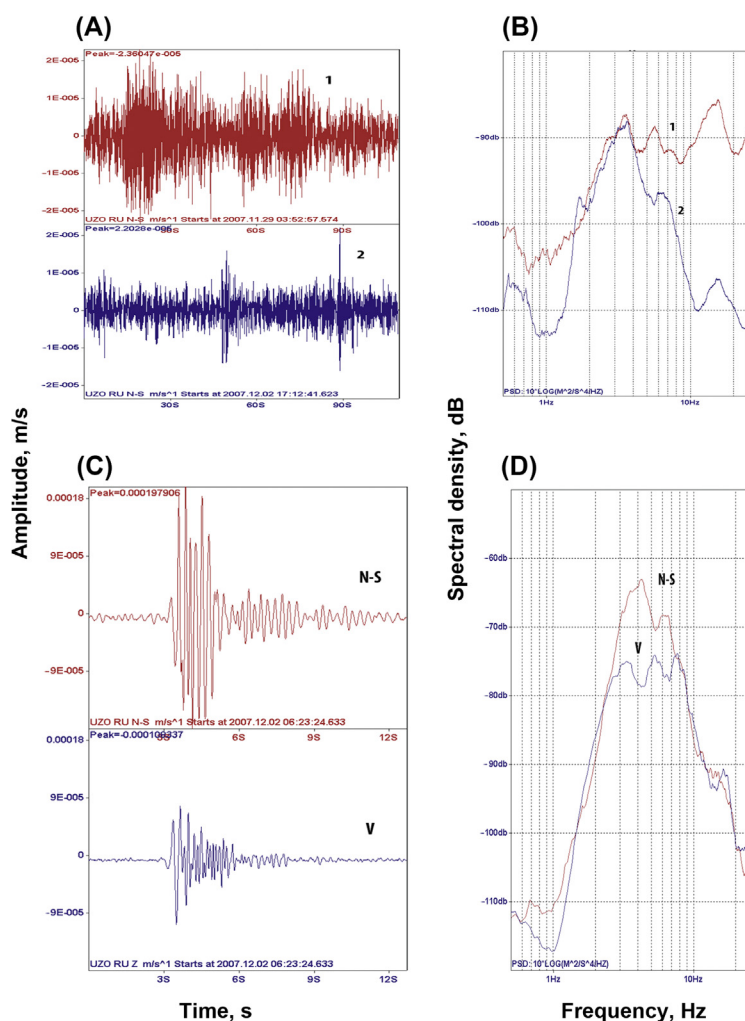


**FIGURE 21.8** Short-period seismic records and spectrograms of 100 Hz sampling data of geothermal tremor associated with the activity of the geothermal system at Caldera Uzon, Russia (seismic station was installed at a distance of about 200 m to the north of the boiling mudpot and hot spring). (A, B) Geothermal tremors recorded during the windy (1) and calm (2) weather, respectively, and their power spectra; (C, D) two-component (N–S, North–South; V, vertical) seismic records of local high-amplitude pulses and their power spectra. After Kugaenko, Y., Saltykov, V.A., Simitsyn, V.I., 2009. *Reconnaissance seismic observations in the Valley of the geysers and Uzon caldera, December 2007. Proceedings of the Annual Conference of Day Volcanology. Institute of Volcanology and Seismology, Petropavlovsk-Kamchatsky, 2009, pp. 173–181 (in Russian).*

Broadband seismic observations, carried out during April 1997 by Ohminato (2006) showed the presence of a continuous geothermal tremor (Fig. 21.9). In the periodic seismic signals observed at Satsuma-Iwojima,

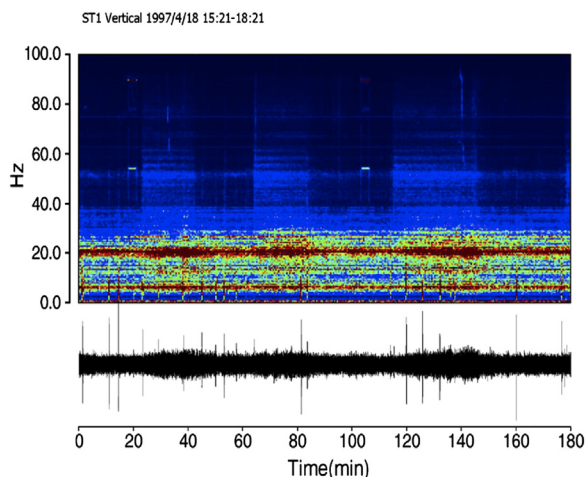


**FIGURE 21.9** Seismic record (below) and spectrogram (above) of 200-Hz sampling data of geothermal tremor associated with the activity of geothermal system at Satsuma-Iwojima, Japan, volcanic island (broadband seismic station was installed at a distance of about 200 m from the crater). From Ohminato, T., 2006. Characteristics and source modeling of broadband seismic signals associated with the hydrothermal system at Satsuma-Iwojima volcano, Japan. *Journal of Volcanology and Geothermal Research* 158, 467–490.

frequency peaks and also frequency-dependent amplitude modulation were also observed. Two predominant groups of frequencies: around 6 Hz and around 20 Hz were identified at the ST1 station, shown in Fig. 21.9. The 20-Hz peak was common for the other three broadband stations installed around the crater but the position of the first frequency peak varied from station to station within 6–12 Hz range. Ohminato (2006) considers that the common 20-Hz peaks can be interpreted as a source effect.

### 21.2.3 Intermittent Heat Discharge

This type of heat discharge is represented by the geyser activity, the phenomena when water and steam spurt from the surface intermittently in a high-reaching fountain. The geyser eruptions are usually accompanied by seismic signals. These seismic signatures vary from one geyser to another. There are two classes of geyser eruptions: fountain and columnar. The fountain eruptions of geysers represent a series of intermittent bursts of bubbles of steam through a surface pool of water. Columnar eruptions of geysers occur as narrow vertical or subvertical columns of ejected water and steam. This type of geyser does not have a surface pool of water (Nicholls and Rinehart, 1967; Kieffer, 1984). The typical seismic signals of the geyser eruptions are described in the following sections.

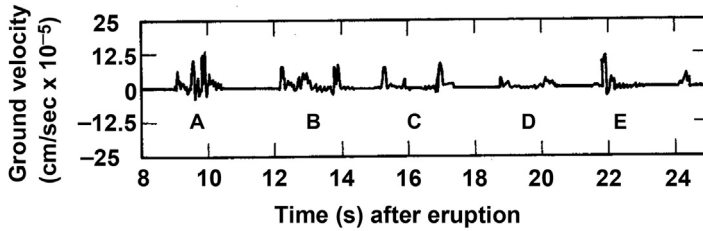


**FIGURE 21.10** Short-period seismic record of the seismic signals following the July 7, 1967 eruption of Strokkur geyser, Iceland. The seismic station was installed at the rim of geyser. From Rinehart, J.S., 1968. *Seismic signatures of some Icelandic geysers*. *Journal of Geophysical Research* 73, 4609–4614.

### 21.2.3.1 Fountain Geyser Strokkur, Iceland

Strokkur's basin is about 12 m in diameter. Its central tube, about 30 cm in diameter, has been built up almost to the surface of the water by deposition of geyserite. The geyser erupts every 10–15 minutes. When superheated water reaches the surface, the water changes suddenly to steam, and the explosion hurls a small quantity of water to a height of 20 m or more. The eruption is short and continues for a few seconds (Rinehart, 1968).

Figure 21.10 shows the seismic signals recorded a few seconds after the beginning of an eruption of the geyser. Strokkur's seismicity is characterized by a series of short, 1–2 second duration, bursts of oscillations that follow the eruption and are separated by 10–20 seconds. The successive seismic bursts are not identical. The first bursts in the series are audible at the surface (Rinehart, 1968). Rinehart (1968) considers that the origin of these seismic signals is supposedly associated with sudden refilling of subterranean cavities, which produces a water hammer effect.

### 21.2.3.2 Columnar Geyser Old Faithful, Yellowstone

The surface expression of this geyser is a 4-m-high, 60-m wide mound with an opening roughly  $2\text{ m} \times 1\text{ m}$  at the top. The geyser's eruptions are 2–5-minute long with the interval between them ranging from 30 to 100 minutes.

Figure 21.11A shows a broadband seismic record of a sequence of the two explosions of the geyser occurring during a 2-hour interval. The seismic signals, preceding the geyser eruption, begin about 70 minutes before an eruption and may be divided into three parts: initial low-amplitude vibrations, high-amplitude vibrations, and preliminary low-amplitude vibrations beginning about 12–15 minutes just before the geyser eruption. The eruptions (shown by arrows with E in Fig. 21.11A) were followed by short seismic signals of about 2-minute duration. Fig. 21.11B gives a detailed 1-minute sequence of the preceding seismic activity. It is seen that this sequence consists of numerous distinct events. Fig. 21.11C shows the three-component records of one of these distinct events together with

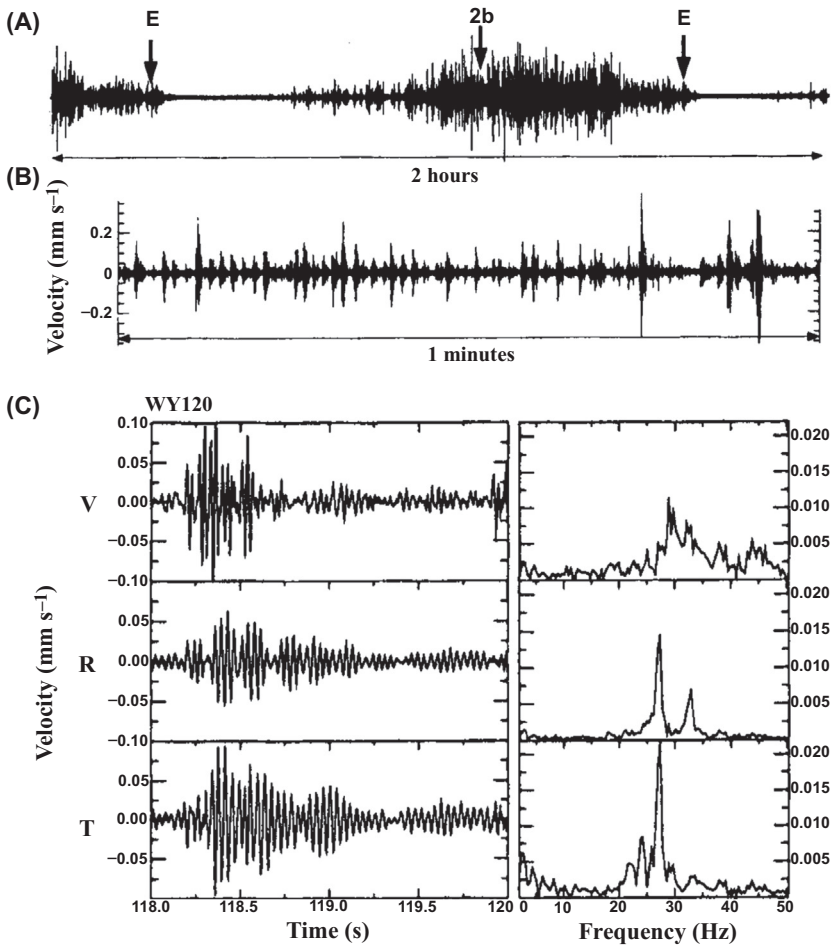


**FIGURE 21.11** Broadband seismic records associated with the 1992 eruptions of Old Faithful Geyser, USA. (A) The 2-h sequence of the seismic signals associated with two explosions (marked by E) of the geyser and recorded at a distance of about 40 m from the crater; (B) the detailed 1-min seismic record of the seismic signals of a sequence of single events preceding the eruption of geyser (this time interval is marked by 2b in A); (C) three-component seismic record (left) and corresponding Fourier spectra (right) of a single event similar to shown in (B) and recorded at a distance of about 30 m from the crater of geyser. *After Kedar, S., Sturtevant, B., Kanamori, H., 1996. The origin of harmonic tremor at Old Faithful Geyser. Nature 379, 708–711.*

corresponding Fourier spectra. The frequency range of the seismic pulses lies between 25 and 35 Hz (Kedar et al., 1996).

An obvious source of the seismic signals associated with Old Faithful Geyser eruptions is supposed to be the repeated collapse of transient steam



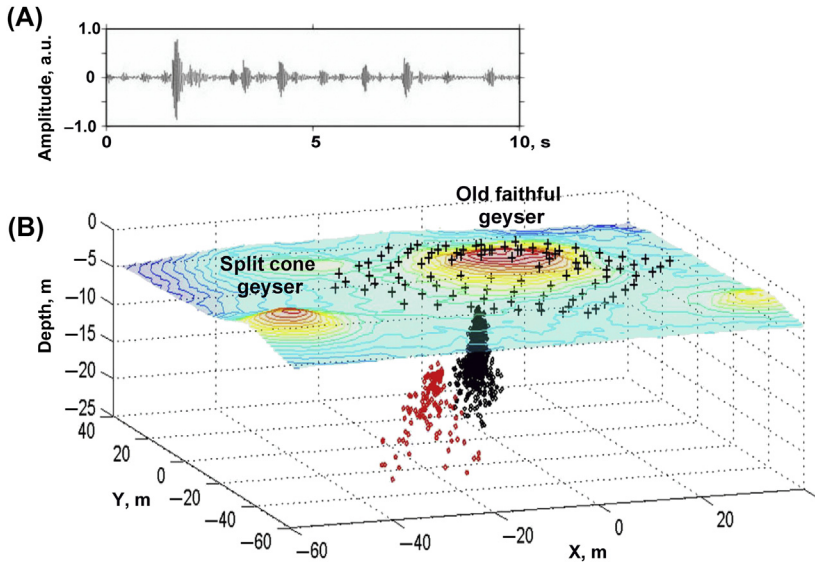


**FIGURE 21.12** A 10-min-long record of tremor at geophone (A) and the 3D representation of the hydrothermal tremor sources at Old Faithful Geyser (OFG) (B). Elevation model (with 1 m interval in color) and location of the 96 vertical geophones (+), with reference to OFG vent are shown in (B). The seismic sources within the geyser conduit and its vertical downward extension are shown with black dots and the sources located within a lateral, cavity-shaped structure are shown with red dots. The unit (a.u.) means arbitrary units. *Modified from Cros, E., Roux, P., Vandemeulebrouck, J., Kedar, S., 2011. Locating hydrothermal acoustic sources at Old Faithful Geyser using matched field processing. Geophysical Journal International 187, 385–393 and Vandemeulebrouck, J., Roux, P., Cros, E., 2013. The plumbing of Old Faithful Geyser revealed by hydrothermal tremor. Geophysical Research Letters 40. <https://doi.org/10.1002/grl.50422>.*

bubbles in the water column (Kieffer, 1984; Kedar et al., 1989). Kieffer (1984) considers that no evidence that rock fracturing is important during eruption cycles.

Vandemeulebrouck et al. (2013) used a seismic array to localize the sources of hydrothermal tremor produced by boiling and cavitation inside the Old Faithful Geyser (Fig. 21.12). The location of seismic sources identified a lateral cavity at 15 m below the surface, on the SW side of the vent, connected to the conduit. The location of the dominant seismic sources at depth discriminates between two structures: the geyser conduit and its vertical downward extension (black dots) and a lateral, cavity-shaped structure (red dots). This reservoir was activated at the beginning of each geyser eruption cycle.

### 21.2.3.3 Columnar Geyser Lone Star, Yellowstone

Lone Star geyser is located about 5 km SSE of the Old Faithful Geyser at an elevation of 2325 m. Lone Star's cone has a diameter of about 5 m at its



base and stands about 3 m above its sinter terrace. There are many small apertures in the upper part of the cone about 1 cm in diameter or less and one main aperture of 20–40 cm in diameter through which the geyser emits most of the liquid water and vapor during eruptions. The eruption cycle is regular and equal to approximately 3 hours; the eruption column reaches a height of 11–12 m (Karlstrom et al., 2013).

Simultaneous measurements of water discharge, acoustic emissions, infrared intensity, and visible and infrared video allowed Karlstrom et al. (2013) to identify four phases in the 3-hour duration eruption cycle of this geyser:

1. a preplay period characterized by a series of 5–10 minute long pulses of steam, small volumes of liquid water discharge,
2. an eruption phase with liquid and steam fountaining, with maximum jet velocities of 16–28 m/s,
3. a posteruption relaxation phase with no discharge from the vents, and
4. a recharge period during which the geyser is quiescent and progressively refills.

Seismicity at Lone Star geyser was dominated by hydrothermal tremor ( $\sim 5\text{--}40$  Hz) attributed to the nucleation and/or collapse of vapor bubbles. Fig. 21.13 demonstrates the variations in the amplitude of hydrothermal

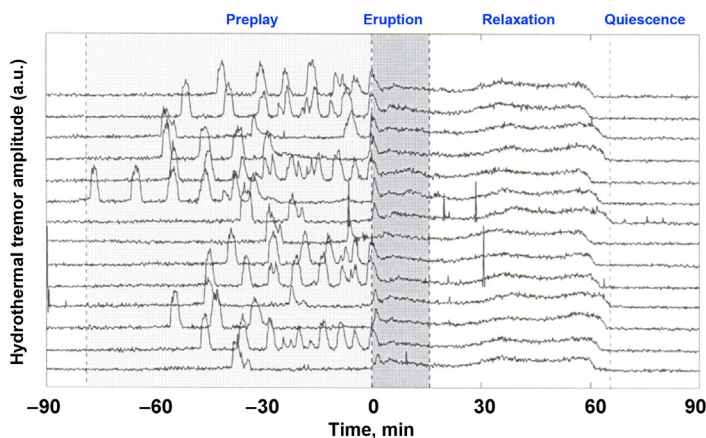


**FIGURE 21.13** Systematic relationship between hydrothermal tremor amplitude and the four-stage eruption cycle at Lone Star geyser. Each horizontal trace represents the high-pass filtered (20 Hz) tremor amplitude for one eruption interval (14 total eruptions shown). Time = 0 marks the start of each eruption. Dashed vertical lines indicate the limits of the different stages; the eruption stage is shaded in dark gray. The unit (a.u.) means arbitrary units. From Vandemeulebrouck, J., Sohn, R.A., Rudolph, M.L., Hurwitz, S., Manga, M., Johnston, M.J. S., Soule, S.A., McPhee, D., Glen, J.M.G., Karlstrom, L., Murphy, F., 2014. Eruptions at Lone Star geyser, Yellowstone National Park, USA: 2. Constraints on subsurface dynamics. *Journal of Geophysical Research Solid Earth* 119. <https://doi.org/10.1002/2014JB011526>.

seismic tremor during the four phases of the eruption cycle that were observed during 14 eruptions (Vandemeulebrouck et al., 2014). The observations demonstrated remarkably consistent characteristics across all eruption cycles. Tremor levels peak at the start of each eruption and then drop to more moderate levels for the duration of the eruptive stage. Immediately following each eruption, tremor levels begin to rise gradually during the relaxation phase. Tremor levels are low during the recharge phase and then exhibit sharp peaks corresponding to each of the preplay pulses leading up to the eruption.

### 21.3 COMPARISON OF THE SEISMIC SIGNALS ASSOCIATED WITH HYDROTHERMAL AND VOLCANIC ACTIVITY

The seismic signals associated with the activity of geothermal systems situated within volcanic environments may be described generally as a continuous low-amplitude geothermal tremor and distinct high-amplitude seismic events. These signals recorded during so-called mild eruptions may be compared with the seismic signals recorded during volcanic eruptions.

Image 2.6 from Chapter 2 showed the variety of the seismic signals that were recorded during the 1998–2011 effusive–explosive eruption of andesitic Volcán de Colima, Mexico and corresponding Fourier spectra. These volcanic earthquakes were associated with explosive and effusive activity of the volcano. Kieffer (1984) and Kedar et al. (1989) pointed out the similarity between the seismic signals associated with the geyser and volcanic activity. Kieffer (1984) compared the seismograms of the banded tremor recorded during the eruption at Karkar volcano, Papua New Guinea, and the seismic records of activity at Old Faithful Geyser, Yellowstone National Park and concluded that short monochromatic high-frequency seismic events recorded before geyser eruption resemble a harmonic volcanic tremor. Kedar et al. (1996) showed that the seismic activity between eruptions at Old Faithful Geyser is composed of a superposition of discrete events originated by pressure pulses inside Old Faithful, in analogy to volcanic long-period events, which are considered the building blocks of continuous volcanic tremor.

Figures 21.14 and 21.15 compare the seismic records and spectra of geothermal tremor and distinct events recorded at the Uzon geothermal system, Kamchatka (Kugaenko et al., 2009) with the seismic records and spectra of volcanic tremor and microearthquakes which often are defined as long-period events (Varley et al., 2010). It is seen that the seismic records of geothermal tremor and volcanic tremor are really similar (Fig. 21.14A and C). At the same time, the spectral characteristics of these seismic signals (Fig. 21.14B and D) are different. The geothermal tremor is characterized by a spectral peak between 3 and 4 Hz, while the volcanic tremor is characterized by peaks at lower frequencies (between 0.7 and 1.2 Hz).

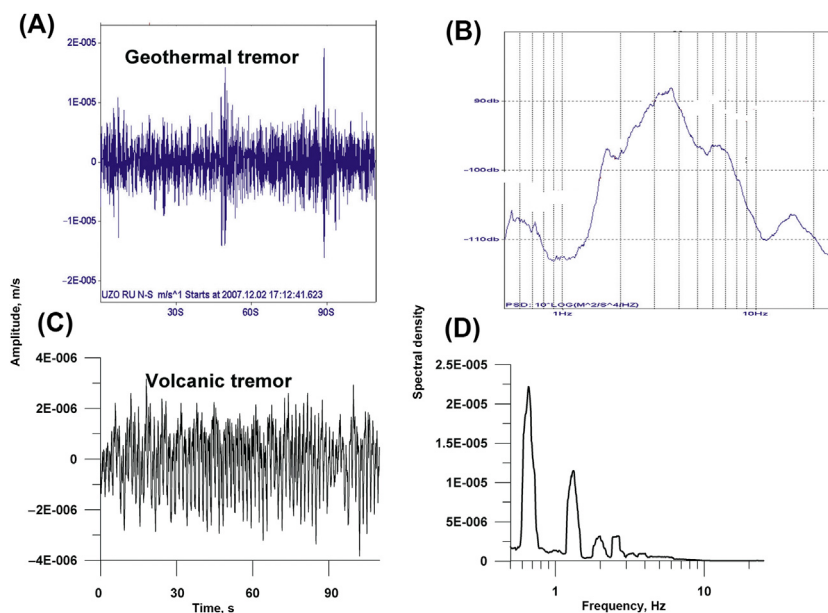


**FIGURE 21.14** Comparison of the seismic records (vertical component) and corresponding spectra of the geothermal tremor recorded at Uzon geothermal system (A, B) and volcanic tremor recorded at Volcán de Colima (C, D). The short-period seismic station at Uzon was installed at a distance of about 200 m to the north of the boiling mudpot and hot spring; the broadband seismic station at Volcán de Colima was installed at a distance of 4 km from the crater. The seismic data of Uzon are taken from Kugaenko, YuA., Saltykov, V.A., Sinitsyn, V.I., 2009. *Reconnaissance seismic observations in the Valley of the geysers and Uzon caldera, December 2007*. In: *Proc. Ann. Conf. of Day Volcanol. Inst Volcanol Seismol, Petropavlovsk-Kamchatsky*, pp. 173–181 (in Russian).

It was selected the seismic record of a microearthquake from the swarm of these events preceding the June 5, 2005 explosion at Volcán de Colima whose waveform was similar to the distinct seismic event recorded at Uzon (Fig. 21.15A and C). In addition, the volcanic event is characterized by lower peak frequencies (Fig. 21.15C; between 1.2 and 2 Hz) compared to the peak frequencies of geothermal events (Fig. 21.15B, between 3 and 8 Hz). So it is possible to say about the similarity of the waveforms of volcanic and geothermal seismic signals but the difference in the spectral contents. This difference in the spectral content indicates the difference in the size of the processes generating similar seismic waveforms, significantly smaller for geothermal activity.

The geothermal systems, situated within the crater zone of the active volcano, may produce seismic signals similar to tornillos (TOR, see Image 2.6 from Chapter 2) as was described for the active cone La Fossa of Vulcano volcano, Aeolian islands (Montalto, 1994; Milluzzo et al., 2010). Tornillos

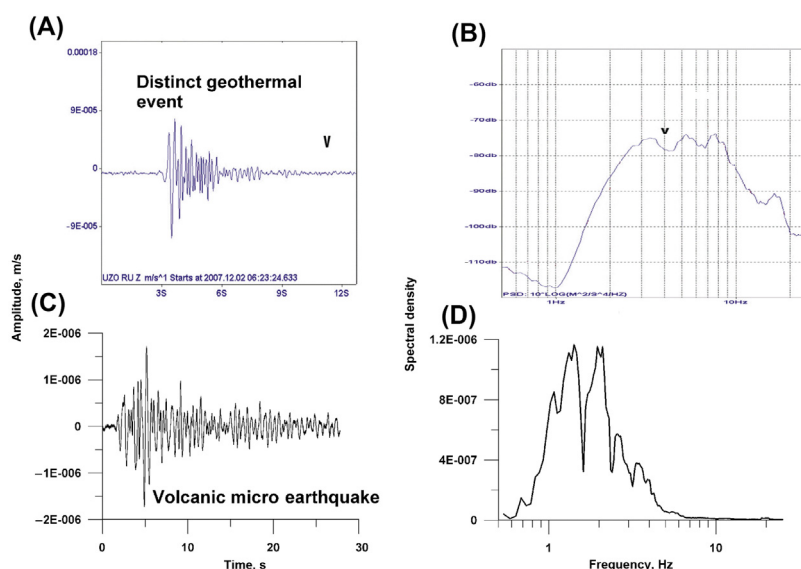


**FIGURE 21.15** Comparison of the seismic records (vertical component) and corresponding spectra of the geothermal distinct event recorded at Uzon geothermal system (A, B) and the microearthquake recorded at Volcán de Colima (C, D). The short-period seismic station at Uzon was installed at a distance of about 200 m to the north of the boiling mudpot and hot spring; the broadband seismic station at Volcán de Colima was installed at a distance of 4 km from the crater. *The seismic data of Uzon are taken from Kugaenko, YuA., Saltykov, V.A., Sinitsyn, V.I., 2009. Reconnaissance seismic observations in the Valley of the geysers and Uzon caldera, December 2007. In: Proc. Ann. Conf. of Day Volcanol. Inst Volcanol Seismol, Petropavlovsk-Kamchatsky, pp. 173–181 (in Russian).*

are sinusoidal waveforms with screw-like envelope profiles that may continue for several minutes and are characterized by a few narrow spectral peaks. They have the properties of volcanic tremors and long-period seismic events; therefore, they are not strange for geothermal systems.

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## **RELEVANT WEBSITES**

GVP, Global Volcanic Program, Smithsonian Institution, <http://www.volcano.si.edu/world/volcano>.

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